W. I. Gardner

COLLEGE TEXT-BOOK OF GEOLOGY

PART I

GEOLOGIC PROCESSES AND THEIR RESULTS

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PREFACE

This work is intended as a textbook for students in the first and second years in college, but has been shaped so as to be serviceable to anyone interested in the principles and products of geologic processes. Each volume contains matter appropriate for a course one quarter or one semester in length. The revision has not only brought the subjects up to date but given them what we think is an improved logical and teachable form. Stress is placed more on processes than descriptions. Our concept of nature's method is given first and the outcome is given later.

The chapter on materials has been so subdivided that each kind is treated as the product of the chief process involved in its formation. A further feature is the introduction of only such material as is considered essential to a sound understanding of the process. The book is intended to give the student so firm a grasp of each process treated that he can take part in constructive class discussion.

Throughout the work an effort has been made not only to set forth the existing state of knowledge, but to present it so that the student will be introduced to the methods and spirit of the science. To this end the working methods of the geologist have been implied as frequently as practicable. To this end also there has been frankness of statement relative to the limitation of knowledge and the degree of trustworthiness of tentative conclusions.

The theoretical and interpretative elements which enter into the general conceptions of geology have been freely used, because they are an essential part of the evolution of the science, because they often help to clear conceptions, and because they stimulate thought. The aim has been, however, to characterize hypothetical elements as such, and to avoid confusing interpretations with the statements of fact or with established doctrines. Especial care has been taken to recognize the uncertain nature of such interpretations as are dependent on unverified hypotheses, especially if this dependence is likely to be overlooked.

Laboratory work coördinated with the class work is considered an important part of the course; but since methods vary widely with individual instructors and institutions, this phase of the work is left to the initiative of the instructor. A pamphlet, covering such work as is conducted by the authors in their classes, may be secured from the publishers.

Field work is likewise considered of major importance; but since opportunities for field work vary from place to place, no adequate consideration can be given it in a textbook. In general, however, we feel that for the average college student "sight-seeing" trips are almost worthless. The benefit of field work, in the opinion of the authors, is derived mainly from the raising of concrete problems on the ground, and mayhap their solution then and there by the students. If not solved at once, hypotheses to be tested as the trip progresses are to be devised, further data gathered, and the subject kept under study. The background of class work should be made the basis for field interpretations and the reverse. Field observations enrich and give concreteness to class discussions.

The revisers wish to express thanks to their many colleagues in the science who have given suggestions; to the United States Geological Survey, the United States Forestry Service, the United States Army Air Service, and the Swiss Air Service for photographs; but particularly to the senior author of the original book who has gone over the manuscript and made important suggestions and additions.

University of Chicago R. T. C.
April 1, 1927 P. M.

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COLLEGE GEOLOGY

GEOLOGIC PROCESSES
AND THEIR RESULTS

CHAPTER I

PRELIMINARY OUTLINE

Geology is Essentially a History of the Earth and its Inhabitants. It treats of rocks and of the agencies and processes which have been involved in their formation, and from the rocks and their structures, it attempts to make out the various stages through which the earth and the living things which have dwelt upon it have passed. It brings under consideration also certain phases of other sciences, particularly astronomy, physics, chemistry, zoölogy, and botany, and is therefore one of the broadest of the sciences.

Subdivisions. So broad a science has many subdivisions. That phase which treats of the outer relations of the earth is *Cosmic* or *Astronomic Geology;* that which treats of the composition of the solid portion is *Petrology*, the science of rocks; that phase which deals with the structural arrangement of the rocks is *Tectonic* or *Structural Geology;* that which deals with the forces involved in geologic processes is *Dynamic Geology;* that which treats of the face of the earth, or topographic form, is *Physiographic Geology;* that which concerns itself with the fossils that have been preserved in the rocks, and with the faunas and floras that have lived in the past, constitutes *Paleontologic Geology* or *Paleontology;* the treatment of the succession of geologic events is *Historical Geology*, which finds its records chiefly in the succession of beds of rock formed in the progress of the ages.

Besides these general subdivisions, there are special applications of geologic knowledge which give rise to other terms: *Economic Geology* is concerned with the industrial and commercial applications of geologic knowledge, and

Mining Geology, which is a sub-section of Economic Geology, with the application of geologic facts and principles to mining. Other similar subdivisions might be mentioned.

Dominant Processes. Three sets of processes, now in operation, have given rise to the surface features of the earth. Many parts of the land, especially along coasts, are known to be slowly sinking relative to sea-level, while other parts are known to be rising. The fact that sediments originally deposited beneath the sea now exist in many places at great elevations, together with the fact that certain areas which were once land are now beneath the sea, proves that similar changes have taken place in the past. Earthquakes also are well-known forms of movement. All such movements of the solid part of the earth, the *lithosphere*, are designated by the term *diastrophism.* Much lava issues from openings in the earth's surface, and much solidifies before reaching the surface. All movements of lava are designated by the term *vulcanism.* Rain, rivers, glaciers, wind, waves, and ground-water are constantly wearing down the higher parts of the lithosphere. This process is *degradation.* The material worn away is deposited on the lower parts of the lithosphere, either on the land, in lakes, or in the sea. This depositing or building up process is *aggradation.* These two processes which tend to bring the surface of the lithosphere to a common level constitute *gradation.*

THE EARTH IN THE SOLAR SYSTEM

The Earth as a Planet. Though supremely important to us, the earth is but one of the minor planets which revolve about the sun. Of the eight planets, four — Jupiter, Saturn, Uranus, and Neptune — are much larger than the earth, while three — Mercury, Venus, and Mars — are smaller. There are a host of planetoids but altogether they do not equal the mass of the smallest planet. The average mass of the planets is more than fifty times that of the earth, and

Jupiter, the largest, has more than three hundred times the mass of the earth. The earth's position is in no sense distinguished. It is neither the outer nor the inner, nor even the middle planet. Even in the inner group of four to which it belongs, it is neither the outermost nor the innermost member, though in this group it is the largest. Its average distance from the sun is about 92,900,000 miles, and its period of revolution is 365¼ days, a period longer than that of any of the inner planets, and shorter than that of any of the outer ones. Its period of rotation is not very different from that of Mars, but is much longer than that of the larger planets whose periods of rotation are known. The plane of the earth's revolution approximates the planes of revolution of all the other planets. The orbit of the earth, like the orbits of the other planets, is an ellipse.

Its Satellite. The earth is peculiar in having one large satellite, which has a mass $\frac{1}{81}$ of its own. Several satellites of the larger planets are larger than the moon; but no other is $\frac{1}{81}$ of the size of the planet about which it revolves. There is little doubt that the moon has played an important part in the history of the earth. It is the chief cause of oceanic tides, and the tides are effective in the wear of ocean shores and in the distribution of marine sediments.

Dependence on the Sun. By far the most important external relation of the earth is its dependence on the sun, of which it is a mere satellite. Its mass is less than one-three-hundred-thousandth that of the sun, upon which it depends for nearly all its heat and light. A little heat and light are derived from other bodies, and an important source of energy is found in the interior of the earth itself; yet all of these are so far subordinate to the great flood of energy which comes from the sun that they are quite inconsequential so far as surface processes are concerned. The dependence of the earth on the sun has been intimate throughout its past history, and its future is locked up with the destiny of that great luminary. Geology in its broadest phases can therefore scarcely be separated from the study of the sun; but this falls in the province of the astronomer rather than the geologist.

THE GRAND DIVISIONS OF THE EARTH

The Constitution of the Earth. The materials of the earth fall into three grand divisions: (1) the *atmosphere*, (2) the *hydrosphere* (water sphere), and (3) the *lithosphere* (rock sphere).

1. THE ATMOSPHERE

Principal Constituents. The atmosphere in its better-known lower part is largely composed of a mixture of two gases — nitrogen about 79 % and oxygen about 21 %.

Minor Constituents. Though small in amount compared with the principal constituents, *water vapor* and *carbon dioxide* are very important. The former varies from place to place at the same time and from time to time at the same place; the amount in the air at one time is but a fraction of 1 % of the mass of the atmosphere. The latter, composing about .03 % of the atmosphere, is practically constant from day to day and year to year. Argon, an inert gas, amounts to .94 %, and also traces of several rare gases are present.

Impurities. Some gases from volcanic vents, from decay of organic matter, and from various chemical processes of manufacture, as well as the solid particles of dust, are looked on as impurities rather than integral parts of the atmosphere. Dust is of importance because it affects the temperature of the air and facilitates the condensation of moisture.

Mass and Extent. The total mass of the atmosphere is estimated to be 1/1,200,000 of the mass of the earth. It exerts a pressure of about fifteen pounds per square inch at sea-level. Its density decreases upward, but there is still enough atmosphere at a height of 200 miles to render meteors incandescent from friction. There is no direct evidence of atmosphere above a few hundred miles, but there are theoretical grounds for believing that it extends out very much farther.

Geologic Activity. The atmosphere is the most mobile and active of the three great subdivisions of the earth. Its direct and indirect effects on water and rocks are so great that it must be regarded as one of the important agents of change. It acts chemically upon the rock substance of the earth, causing hardening of the rock in some cases, but in more cases causing disintegration, by means of which rock is reduced to soil material, and prepared for removal by winds and waters. When in motion, the atmosphere acts mechanically on the surface of the land, transporting dust and sand. Its greatest function, however, is in furnishing the conditions for water action. Rains, streams, ground-water and glaciers are dependent in one way or another on the atmosphere. On the ocean, too, wave action is caused largely by wind. Streams and waves, which are the most familiar agents of geologic change, are therefore to be credited almost as much to the atmosphere as to the hydrosphere.

A Thermal Blanket. The atmosphere is a thermal blanket to the rest of the earth. If it were absent the heat of the sun would reach the surface with much greater intensity than now, and it would be radiated back from the surface almost as rapidly as received. During the night, there would be a degree of cold far greater than that now known to any part of the earth. In passing through the atmosphere, certain portions of the radiant energy of the sun are absorbed. Of the remainder which reaches the surface of the earth, a part is transformed into vibrations of longer wave length which are then more effectively retained by the atmosphere. The air thus distributes and equalizes temperature. The two constituents of the atmosphere which are most efficient in this work are water vapor and carbon dioxide, and the climate of the earth is believed to have been very greatly affected by the varying amounts of these constituents in the atmosphere, as well as by the variation in the total mass of the atmosphere.

The function of the atmosphere in sustaining life and promoting all that depends on life is obvious.

2. THE HYDROSPHERE

The water of the oceans, lakes, and rivers, together with that in the pores and fissures below the surface of the land, makes up the *hydrosphere*. All other waters of the earth, however, are small in amount compared with the great mass of the oceans. Were the higher parts of the lithosphere planed down and the lower parts filled up till it was perfectly spheroidal in shape, a universal ocean nearly two miles deep would cover the globe. Owing to the irregularities of the lithosphere the water of the oceans is gathered into great basins and occupies about three-fourths (72%) of the earth's surface. These basins are all connected so that anything which changes the level in one, changes the level in all.

Oceanic Dimensions. The surface of the ocean is estimated at about 143,000,000 square miles. Beneath about 20% of this area, the bottom sinks to depths between 6,000 and 12,000 feet; under about 53% it sinks to depths between 12,000 and 18,000 feet; and under 4% it ranges from 18,000 feet down to about 30,000. The average depth of the ocean is about 2½ miles.

Geologic Activity. Of all geologic agents, water is the most obvious and apparently the most effective, though its effectiveness is modified by many conditions, especially the relief of the land, and the temperature. Through the agency of rainfall, surface streams, underground waters, and waves, the hydrosphere is constantly modifying the surface of the lithosphere, carrying sediment from the land and depositing it in various basins. The hydrosphere thereby becomes the great agency for degrading the land and building up the basin bottoms. It is, therefore, both destructive and constructive in its action. The beds of sediments which have been laid down in it follow one another in orderly succession, each later one lying above an earlier, in this way forming a time record. And as relics of the life of each age have become imbedded in the sediments, they now furnish the means of following the history of life from age to age. The historical record of geology is therefore very largely dependent upon the fact that the waters have buried, in systematic order, relics of the life of the long past.

3. THE LITHOSPHERE

The atmosphere and the hydrosphere are envelopes or shells surrounding and to a limited extent penetrating the solid part of the earth, *the lithosphere.* The lithosphere is an oblate spheroid in shape, being flattened by its rotation, till its polar diameter, 7,899.7 miles, is 26.8 miles shorter than the equatorial diameter. The polar flattening and equatorial bulging result in its equatorial circumference, 24,902 miles, being 42 miles greater than the meridional circumference. The surface area of the lithosphere is about 197,000,000 square miles of which 143,000,000 square miles lie below the ocean and 54,000,000 square miles are land.

Irregularities. The existence of land is due to the fact that the surface of the lithosphere has pronounced irregularities. These irregularities, or relief features, may be grouped into three orders:

(*a*) *Relief Features of the First Order.* The most important irregularity of the surface of the lithosphere is the existence of the great depressed areas, the *ocean basins,* and the broad elevated areas, the *continental platforms* (Fig. 1). The ocean basins and the continental platforms are the *relief features of the first order.* The continental platforms (64,000,000 square miles) are somewhat larger than the continents (54,000,000 square miles), while the ocean basins (133,000,000 square miles) are somewhat smaller than the oceans (143,000,000 square miles). The discrepancy results from the fact that there is more water in the oceans than the true basins can hold, so that some of it spreads upon the low margins of the continental platforms to form *epicontinental* (upon the continent) *seas*. The marginal portions of the continental platforms slightly submerged beneath oceanic waters at the present time are the *continental shelves*, aggregating about 10,000,000 square miles.

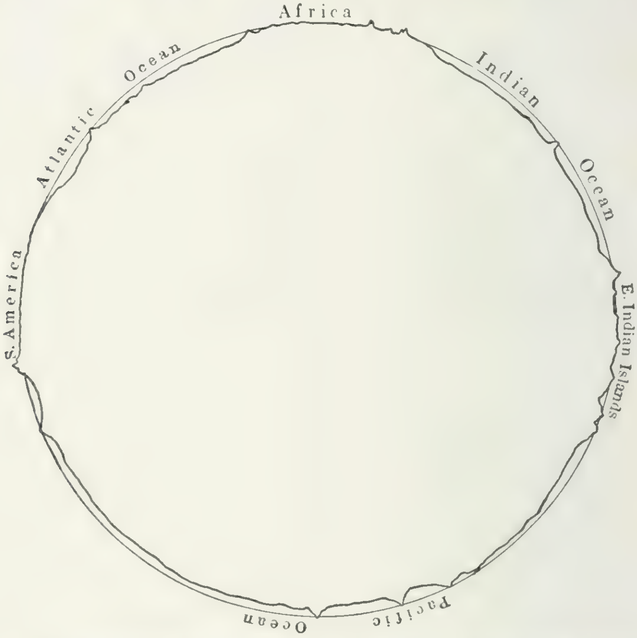

Fig. 1. — Diagrammatic section of the earth about the equator, showing the elevated segments (continents) and the depressed segments (ocean basins). Vertical scale X40. (Based on section in Stanford's Atlas of Universal Geography.)

Since the average depth of the ocean is about 2½ miles while the average height of the land above sea-level is about ½ mile, what might be called an average relief of the lithosphere amounts to about 3 miles. If drawn to scale, this relief would be one inch on the surface of a globe 110 feet in radius.

It is customary to look upon the protrusions of the continents as the great features of the earth's surface, but in reality the oceanic depressions are the master features. Both in breadth and depth they much exceed the continental protrusions and, if the earth be regarded as a shrunken body, the settling of the ocean bottoms has doubtless been the greatest diastrophic movement.

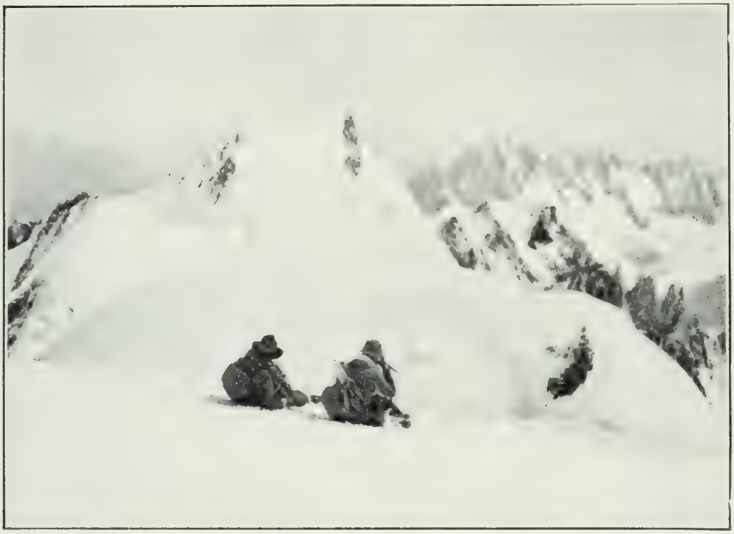

Fig. 2. — The chain of French Alps showing the small summit area. From near the summit of Mont Blanc. (R. T. Chamberlin.)

(*b*) *Relief Features of the Second Order.* The continental platforms are diversified by three major subdivisions, *plains, plateaus* and *mountains.* Plains are the lowlands of the earth. They cannot be defined in terms of altitude, but if high they are invariably low with respect to neighboring features. Coastal plains border many continents, while interior plains exist in many parts of the earth, at least partly separated from the sea by highlands. Plateaus are the highlands of the earth which have considerable summit area. They stand conspicuously above their surroundings on at least one side. Mountains are the highlands of the earth which have but slight summit area (Fig. 2). Mountains also cannot be defined in altitude, but they must stand conspicuously above their surroundings (Fig. 3).

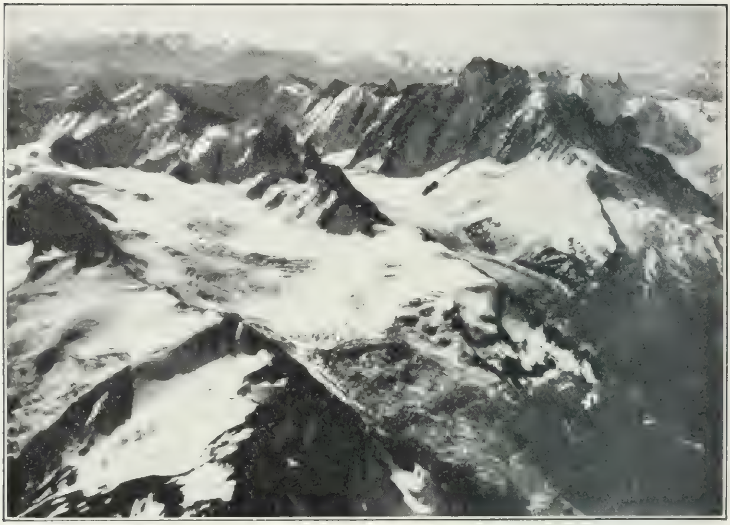

Fig. 3. — Airplane photograph of the Alps. (Swiss Air Serv.)

(*c*) *Relief Features of the Third Order.* Features of lesser size, such as ridges, hills, valleys, and basins, are those of the third order.

The Interior. Until recently little has been known about the interior. But the transmission of earthquake waves, and the periods of body tides and polar oscillations are rapidly adding evidence that will soon make the chief nature of the interior fairly certain. The evidence at hand shows that the earth is an *elastic solid* with rigidity greater than that of steel. From the weight of the earth (its specific gravity as a whole is 5.53 and the specific gravity of its outer portion is about 2.7) it is inferred that its interior is much more dense than its surface. From the phenomena of volcanoes, and from observation of temperature in deep borings, it is inferred that its interior is very hot. Further inferences concerning its character call for less simple statement, and will be discussed later.

The processes of gradation, diastrophism and vulcanism will be taken up in succession. But before entering on these considerations, a brief summary of the materials upon which they act is needed.

MATERIALS OF THE LITHOSPHERE

The Mantle Rock. The surface of most land is covered with vegetation. Beneath this vegetation there is loose material consisting of clay, sand and gravel, and broken rock, known collectively as *mantle rock*. This layer varies from a few inches to hundreds of feet in thickness. The upper part of the mantle rock is commonly called the soil.

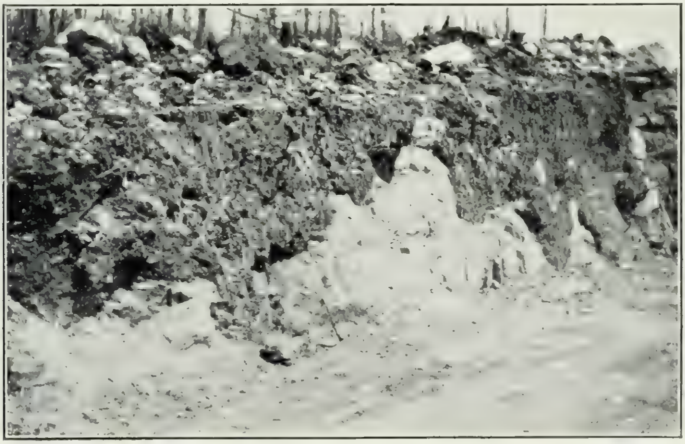

Fig. 4. — Soil and subsoil, arising from the decay of limestone, resting on the uneven surface of the rock beneath. Southeastern Missouri. (Buckley.)

In places the mantle rock grades so imperceptibly into the underlying bed-rock as to show that this loose material has been derived from the decay of the solid rock beneath (Fig. 4). Mantle rock of this type is evidently *local* in origin. In other areas the contrast and contacts between mantle rock and solid rock are distinct and sharp. The mantle rock in such cases commonly contains materials that could not have been derived from the rock below (Fig. 5). They have been *transported* from some other place by agents such as streams, waves, winds, or glaciers.

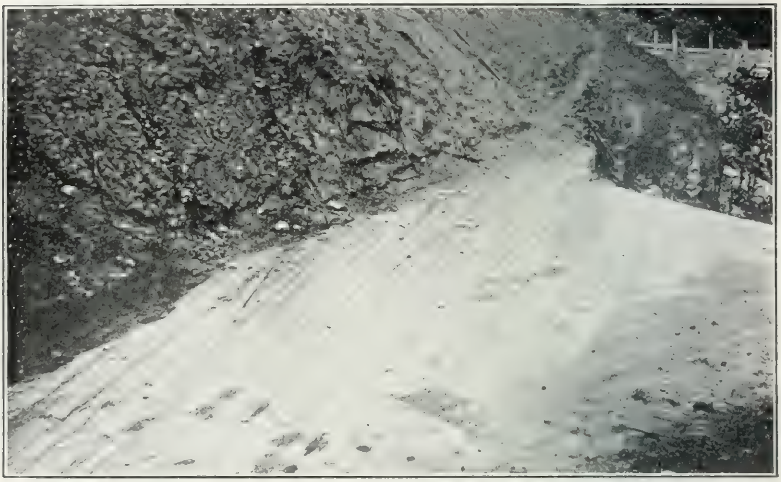

Fig. 5. — Transported mantle rock lying on undecayed bed rock. Kingston, Des Moines County, Iowa.

The Solid Rock. Mantle rock is absent in some places, and there the *solid rock* appears. It is common on the slopes of steep-sided valleys and mountains, and on the slopes of cliffs which face seas or lakes, and is frequently seen in the channels of swift streams, especially where there are falls or rapids. In all lands inhabited by civilized peoples, there are numerous wells and other excavations, ranging from a few feet to several hundred feet in depth, which encounter solid rock. It may, therefore, be accepted as a fact that the upper surface of the solid rock is nowhere very far below ground.

Varieties of Solid Rock. The rocks of the earth may be grouped into three great classes, — *sedimentary*, *igneous*, and *metamorphic*.

1. *Sedimentary Rocks*, as their name implies, are made of mud, sand, and gravel (sediments), such as are being deposited at present by rivers, lakes, and seas. They appear in most places in layers, or beds, varying from a few inches to many feet in thickness. Because of this bedded arrangement (stratification) they are also called *stratified* rocks (Fig. 6). The stratification of sedimentary rocks and that of recent sediments is the same, and the markings on the surfaces of the layers, such as ripple-marks, rill-marks, wave-marks, etc., are identical. Furthermore, many of the stratified rocks of the land, like recent sediments of the sea, contain the shells and skeletons of animals, and in places the impressions of plants. Most of the relics of past life (fossils) found in the stratified rocks belong to animals or plants which lived in marine water. Because of their structure, their composition, their distinctive markings, and the remains of life which they contain, it is confidently inferred that most of the stratified rocks which lie beneath the mantle rock of the land were originally laid down in beds beneath the sea, and that the familiar processes of the present time furnish the key to their origin.

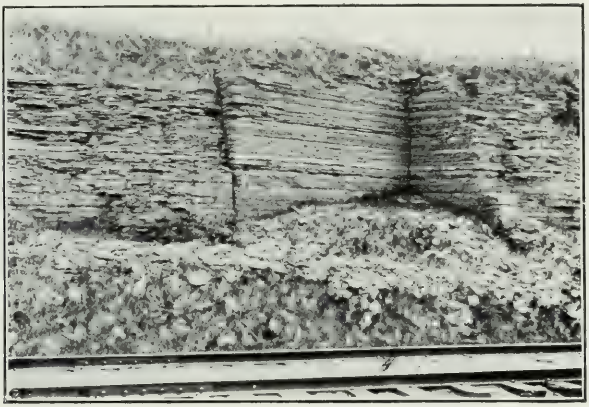

Fig. 6. — Stratified rock. Platteville limestone, Fort Snelling, Minn. (Calvin.)

If rounded pebbles of gravel become cemented into a solid rock by deposition of mineral matter around and among them, a *conglomerate* is formed. If the pieces of rock are not rounded into pebbles, but are angular, a *breccia* is formed. Likewise when grains of sand are consolidated by cementation, sandstone results. And when particles of mud are compacted and cemented they form *shale*. The cementing material, in general, was deposited from solution in the water which at some time overlay, filled, or circulated through the sediments. Conglomerate, sandstone, and shale, made up chiefly of particles derived from the breaking up of other rock, are *clastic* (broken) rocks. It is clear, therefore, that

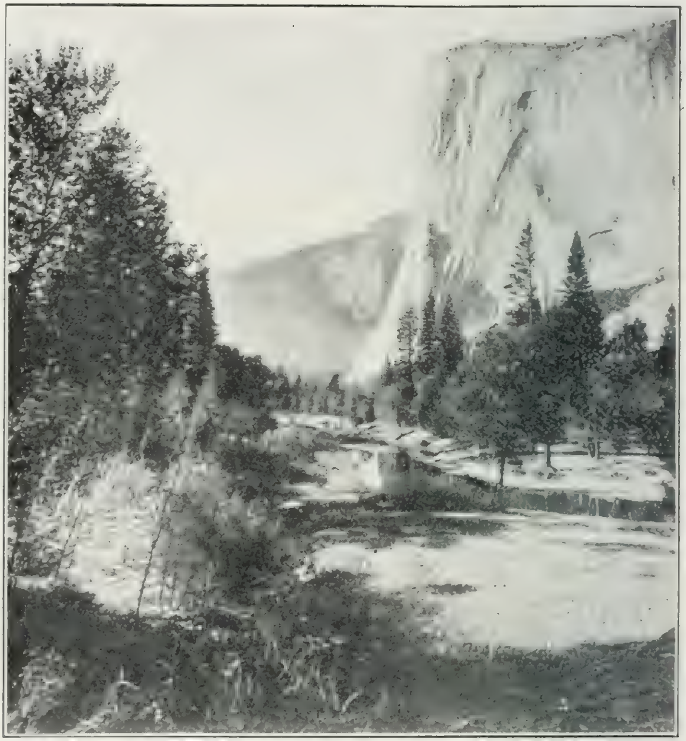

Fig. 7. — Massive rock. El Capitan, Yosemite, Calif. (Courtesy, U. S. Forestry Serv.)

the destruction of one generation of rocks furnishes the material for another and younger generation.

Life exists practically everywhere in the shallow sea. When shell-bearing animals die, their shells, composed predominantly of calcium carbonate ($CaCO_3$), settle to the bottom. If sand or mud is accumulating faster than the shells, a limy sandstone or limy shale results. But if clastic materials are in the minority, the shells, partly ground up by the waves and augmented by chemically precipitated calcium carbonate from the sea water, form a lime mud which, upon solidification, makes *limestone*. After the calcium carbonate has accumulated, the calcium may become partly replaced by magnesium to form a calcium magnesium carbonate ($CaMgCO_3$). Such a rock is known as *dolomite*.

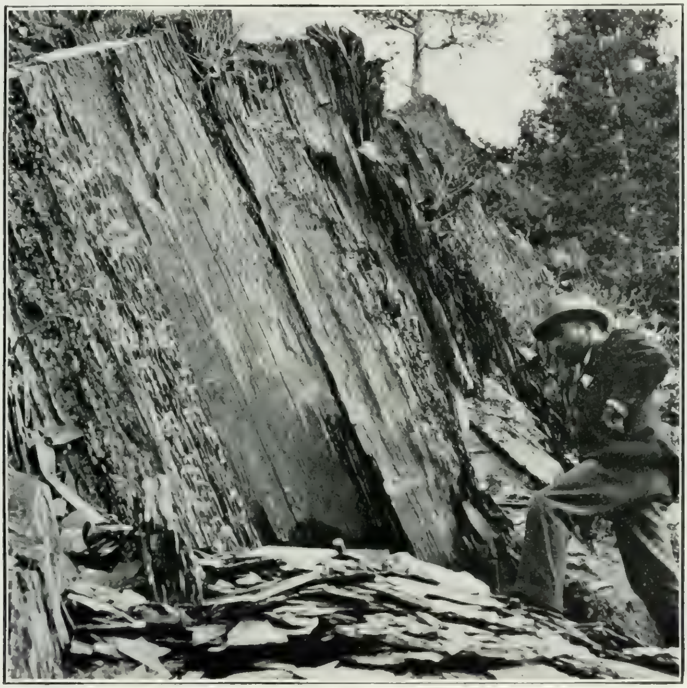

Fig. 8. — Pre-Cambrian fossiliferous slate. Deep Creek Canyon, 16 miles southeast of Townsend, Mont. (Walcott, U. S. Geol. Surv.)

The various kinds of sedimentary rocks are rarely pure. Some mud is normally found in sand, some sand in mud. Likewise calcium carbonate is to be expected in the clastic rocks as well as clastic material in limestone. Sandy shales, shaly limestones, etc., are, therefore, very common.

2. *Igneous Rocks.* Hot liquid rock material issues from vents in the earth's surface. This liquid rock material is *lava.* Such lava solidifies at the surface, but under other conditions much lava solidifies before it reaches the surface, and is exposed to view only after the covering which once overlay it has been eroded away. Lava is of many varieties: some is light colored, some is dark, some is very fluid, while some is more viscous. All the different kinds of rocks formed by the solidification of lava are known as *igneous rocks.* Since they are usually not in beds or strata, they are spoken of as *non-stratified*, or *massive* (Fig. 7).

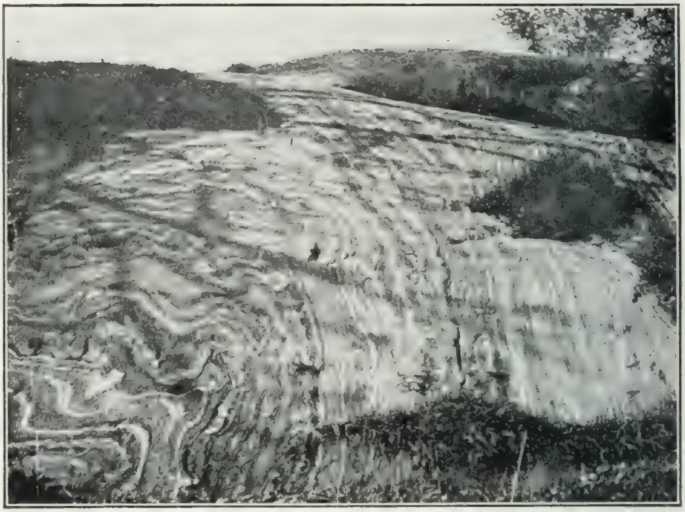

Fig. 9. — Metamorphic rock. (Ells, Can. Geol. Surv.)

3. *Metamorphic Rocks.* These are the rocks, of either sedimentary or igneous origin, which have suffered notable change from their original condition. In some cases this

change has been so great that it is a difficult matter to tell whether the rock was originally igneous or sedimentary. Metamorphism takes place (1) under the influence of great pressure which alters the structure of the rock, in many cases flattening and rearranging the particles to produce cleavage (Fig. 8); (2) through the operation of great heat, causing the minerals to recrystallize and recombine in new forms; and (3) by the dissolving and precipitating action of water, which changes the composition of the rock. Two or more of these agents of metamorphism commonly act together to produce such metamorphic rocks as quartzite, marble, gneiss (Fig. 9), slate, or schist.

REFERENCE LIST OF THE COMMONER ROCKS[1]

Adobe. A fine silty or clayey deposit formed by gentle wash from slopes and subsequent lodgment on flats; especially applied to clayey accumulations in the basins and on the plains of the western dry region.

Agglomerate. An aggregate of irregular, angular, or subangular blocks of varying sizes, usually of volcanic origin; distinguished from conglomerate, in which the constituents are rounded.

Alluvium. Sediment deposited on land by running water.

Amygdaloid. A vesicular igneous rock whose cavities have been filled with minerals deposited from aqueous solutions; the fillings are called *amygdules,* because sometimes almond-like in form.

Andesite. A fine-grained or dense, extrusive igneous rock consisting essentially of the plagioclase feldspar *andesine* (sometimes oligoclase) and pyroxene (or some related ferromagnesian mineral); sometimes cellular, porphyritic, or even glassy.

Anorthosite. A rock consisting mainly of the feldspar labradorite.

Aphanite. An igneous rock whose crystals are so minute as to be indistinguishable to the naked eye; rather a condition of various rocks than the name of any specific rock.

Arenaceous Rocks. Either those which are mainly sand, or those in which sand is a notable accessory.

Argillite. A clayey rock; usually applied to hard varieties only.

Arkose. A sand or sandstone composed in large part of grains of feldspathic minerals derived from disaggregated granite, or similar rocks. Sand, when undefined, is understood to be quartzose.

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[1] These definitions are given, as nearly as practicable, in accordance with present common usage, which is, however, more or less varying and inconsistent.

Augitite. A rock made up mainly of augite.

Basalt. A dark, compact, basic, extrusive igneous rock consisting of a mass of minute crystals sometimes with more or less glassy base, often containing also visible crystals; composed of basic plagioclase and pyroxene, with olivine, magnetite, or titaniferous iron as common accessories; a basic lava in which the crystallization has taken place rapidly; graduates into dolerite and basic andesite.

Bituminous Coal. Common soft coal, intermediate between lignite and anthracite; contains much bituminous matter, i.e., hydrocarbons.

Boulders. Rounded masses of rock, such as those that have been shaped by glaciers, or streams; larger than cobbles.

Breccia. A rock composed of angular fragments, contrasted with conglomerate the fragments of which are rounded.

Calc-sinter (calcareous tufa). A loose cellular deposit of calcium carbonate made by springs; travertine is a better term.

Cannel Coal. A very fine-textured homogeneous bituminous coal, giving off much gas and burning with a candle-like flame.

Chalk. A fine-grained soft rock composed essentially of calcium carbonate derived from minute marine organisms.

Chlorite Schist. A schistose rock in which chlorite is a predominant mineral; usually greenish, whence the name.

Clastic Rock. A sedimentary rock formed from the debris of broken-down rocks; includes conglomerate, sandstone, shale, etc.

Clay. A term commonly applied to any soft, unctuous, adhesive deposit, containing a considerable amount of aluminum silicate; many so-called clays are chiefly silicious silts or loams.

Clay Ironstone. A clayey rock heavily charged with iron oxide, usually limonite; commonly in concretionary form.

Chert. A general term for compact, non-crystalline silica, usually of light color, occurring abundantly in concretionary form as nodules in certain limestones.

Coal. A carbonaceous deposit formed from the preservation of plant material with the loss of much of the volatile matter.

Concretions. Aggregates of rounded outlines formed about a nucleus; the material is various: iron oxide, calcite, silica, etc.

Conglomerate (puddingstone). A rock formed from rounded pebbles; cemented gravel.

Coquina. A rock formed almost wholly of small and broken shells; especially applied to a shell limestone of Florida.

Dacite (quartz-andesite). An andesite with quartz.

Diabase. As used in Europe, an ancient basalt which has undergone alteration.

Diatom Ooze. A soft silicious deposit found on the bottom of the deep sea, made largely or partly of the secretions of diatoms.

Diorite. A coarse-grained intrusive igneous rock usually of dark-greenish or gray color, consisting of acid plagioclase feldspar and hornblende; often speckled from the commingling of light feldspar and dark hornblende.

Dolerite. A term used in Europe for coarse-grained basaltic rocks; includes diabase as used in America.

Dolomite. A magnesian limestone.

Drift. In common American usage, a mixture of clay, sand, gravel, and boulders formed by glacial agencies.

Fanglomerate. The poorly sorted material of alluvial fans.

Felsite. A general term for any light-colored, fine-grained or dense rock composed of feldspar, often with quartz, in which the crystallization is very imperfect or obscure, giving a close-grained texture with conchoidal fracture and flinty aspect.

Fire Clay. A clay capable of standing a high degree of heat. The percentage of iron, calcium, magnesium, and the alkalies must be very low. Often found in association with coal.

Flint. A compact dark chalcedonic or lithoid form of silica; an impure chert.

Fulgurites. Glassy tubes, produced through fusion by lightning in penetrating sand, earth, or rock.

Gabbro. A coarse-crystalline plutonic rock composed of the plagioclase feldspar labradorite (or anorthite), and diallage (or a related ferromagnesian mineral) with magnetite or titaniferous iron as a common accessory.

Gangue. A term applied to the crystalline vein-material in which ores are imbedded.

Ganister. Essentially a quartz silt, pulverized quartz, or silicious fire clay suitable for lining iron furnaces.

Geyserite. The silicious sinter deposited about hot springs.

Gneiss. A banded rock, consisting normally of quartz, feldspar, and mica; the feldspar typically orthoclase.

Granite. A granular crystalline rock composed of quartz, feldspar, and mica; the feldspar typically orthoclase; popularly and properly used for any distinctly granular, crystalline igneous rock.

Graphic Granite. (See pegmatite.)

Greensand. A sand or sandstone containing a notable percentage of grains of glauconite.

Greenstone. A comprehensive term used to designate igneous and metamorphic crystalline rocks of greenish hue and of intricate and often minute crystallization; they are mostly dolerites, diabases, and diorites; a convenient term for field use where the constituents cannot be determined, and for general use when the variety is unimportant.

Graywacke. An argillaceous rock containing in addition to grains of basic silicates, etc., fragments of shale, basalt and other rocks.

Hornblendite. A rock essentially composed of hornblende.

Hornstone. A very compact, silicious rock of hornlike texture, allied to flint; term sometimes also applied to flinty forms of felsite (hornfels).

Hydraulic Limestone. (See waterlime.)

Ironstone. A rock composed largely of iron oxide, usually applied to clayey rocks having a large iron content.

Infusorial Earth (tripolite). An earthy or silt deposit consisting chiefly of the silicious secretions of diatoms.

Itacolumite. A flexible sandstone whose pliability is due to an open arrangement of sand grains which are loosely held together.

Jasper. A reddish variety of chert.

Lapilli. Small fragments of lava ejected from volcanoes; volcanic cinders.

Lignite (brown coal). A soft, brown or black, low-grade, impure coal.

Limestone. A sedimentary rock composed primarily of calcium carbonate, though magnesium sometimes replaces a part of the calcium. (See dolomite and marble.)

Loess. A very fine, porous, silicious silt containing some calcareous material which often collects in nodules (*Löss Kindchen*) or in vertical tubules; characterized by a peculiar competency to stand in vertical walls.

Mantle Rock. (See regolith.)

Marble. Typically a granular crystalline limestone or dolomite produced by metamorphic action; but the term is variously applied to calcareous and even to other rocks that are colored ornamentally and susceptible of polish.

Marl. An earth formed largely of calcium carbonate, usually derived from the disintegration of shells; or the calcareous secretions of plants, notably the stoneworts; term also applied to glauconitic and other fertilizing earths.

Meta-igneous Rock. A metamorphosed igneous rock.

Metamorphic Rock. A rock which has been altered, particularly one which has been rendered crystalline, or recrystallized by heat and pressure in conjunction with water.

Meta-sedimentary Rock. A metamorphosed sedimentary rock.

Microgranite. A very fine-grained granite.

Microlites. Incipient crystals found in glassy lavas; in many cases needle-shaped, or rod-like; occurring singly or in aggregates.

Monzonite. A coarse-grained intrusive igneous rock composed of orthoclase and plagioclase in nearly equal proportions, with ferromagnesian minerals; a rock intermediate between syenite and diorite.

Mudstone. Solidified mud or silt, shale.

Nodules. Concretionary aggregations of rounded form.

Novaculite (hornstone, oilstone). A very fine-grained, hard sandstone or siltstone, used for whetstones.

Obsidian. A typical form of volcanic glass usually of the acidic class.

Onyx. A variety of chalcedonic quartz having colored bands alternating with white; the "Mexican onyx" is a crystalline calcium carbonate, variegated with delicate colors due to iron and manganese.

Oölite. A limestone or dolomite composed of small concretions resembling the roe of fish.

Ooze. A soft deposit of the deep sea; some kinds characterized by microscopic shells from which they are mainly derived; as globigerina ooze, etc.

Orthogneiss. A gneiss produced by metamorphism of an igneous rock.

Paragneiss. A gneiss produced by metamorphism of a sedimentary rock.

Peat. The dark brown or black residuum arising from the partial decomposition and preservation of vegetable tissue in marshes and wet places.

Pegmatite. A term of ill-defined usage applied to dike-rocks whose crystals commonly are very large; often takes on peculiar aspects due to the simultaneous crystallization and mutual inter-growths of the crystals; graphic granite is a distinct type of pegmatite in which quartz and orthoclase crystals grew together along parallel axes so that cross-sections give figures resembling certain Semitic letters.

Peridotite. A very basic, coarse-grained igneous rock composed chiefly of olivine with augite or related ferromagnesian minerals.

Perknite. A group-name for coarse-grained, very basic igneous rocks containing little or no feldspar; includes peridotite and pyroxenite.

Perlite (pearlstone). A form of glassy lava fractured by concentric cracks so that it appears as though made up of small spheroids.

Phonolite (nephelite-trachyte, clinkstone). A compact igneous rock formed of sanidine and nephelite, with accessories; the extrusive equivalent of nephelite syenite.

Phyllite (argillite). A variety of indurated, partly metamorphosed, clay silt in which finely disseminated micaceous scales are abundant and lustrous; intermediate between typical clay slate and mica-schist.

Pitchstone. A dark vitreous, acid, igneous rock of less perfect glassy texture than obsidian, and more resinous and pitch-like.

Plutonic Rocks. Igneous rocks formed deep within the earth under the influence of high heat and pressure; hypogene rocks; distinguished from eruptive rocks formed at the surface.

Porphyry. A rock consisting of large crystals embedded in a fine-grained or dense ground-mass, or of large crystals embedded in smaller ones.

Pumice. A glassy form of lava rendered very vesicular through inflation by steam; vesicles very small.

Pyroclastic Rocks. Fragmental or clastic rocks produced through igneous agencies, embracing volcanic dusts, tuffs, agglomerates, etc.

Pyroxenite. A coarse-grained igneous rock consisting almost entirely of pyroxene.

Quartzite. A sandstone thoroughly cemented by silica.

Regolith. The earthy mantle that covers indurated rocks, chiefly residuary earths; mantle-rock.

Rhyolite. A fine-grained or partly glassy acidic extrusive igneous rock showing flow lines.

Sandstone. Indurated sand usually composed of grains of quartz, but not necessarily so; sometimes contains grains of the various silicates.

Schist. A crystalline rock having a foliated or parallel structure, splitting easily into slabs or flakes, less uniform than those of slate; they are composed mainly of the silicate minerals, with little or no feldspar.

Scoria. Light, cellular fragments of volcanic rock, coarser than pumice.

Septaria. Concretions the interior cracks of which have been filled with calcite or other mineral deposited from solution (Fig. 57).

Serpentine. A rock consisting largely of the mineral serpentine; derived in most cases by alteration from magnesian silicate rocks.

Shale. A more or less laminated rock, consisting of indurated muds, silts, or clays.

Slate. A metamorphosed shale which splits readily along cleavage planes which are generally inclined to the bedding planes.

Soapstone (steatite). A soft unctuous rock, composed mainly of talc, or talc and chlorite.

Stalactites. Pendant icicle-like forms of calcium carbonate deposited from dripping water.

Stalagmites. The complement of stalactites formed by calcareous waters dripping upon the floors of caverns.

Steatite. (See soapstone.)

Syenite. A granitoid plutonic rock composed of orthoclase and hornblende, or other ferromagnesian mineral without visible quartz.

Till (boulder clay). A stony or bouldery clay, or rock rubbish formed by glaciers; physically and lithically heterogeneous.

Trachyte. A compact, usually porphyritic, extrusive igneous rock, consisting mainly of sanidine, biotite and hornblende.

Trap. A general term for extrusive igneous rocks of the darker basaltic types.

Travertine. A limestone deposited from calcareous waters, chiefly springs; usually soft and cellular, and hence also called calcareous tufa; calc sinter.

Tuff. A term including certain porous granular or cellular rocks of diverse origins; the volcanic tuffs embrace the finer kinds of pyroclastic detritus, as more or less indurated dust, cinders, etc.

Waterlime. An impure argillaceous limestone which, when burned, makes hydraulic cement.

CHAPTER II

THE ATMOSPHERE

The atmosphere is one of the three great subdivisions of the earth, and as a factor in geology it takes its place beside the hydrosphere and the lithosphere. It has played a part in the history of the earth comparable to that of water, though its record is less clear. Unsubstantial as the atmosphere seems when contrasted with the liquid and solid portions of the earth, its extreme mobility and its chemical activity make it an agent of importance in geological history. It plays and has played a direct part as (1) a mechanical and (2) a chemical agent, and it serves and has long served an indirect function in furnishing the conditions under which (3) the sun produces its temperature effects, and (4) evaporation and precipitation take place. In addition, the atmosphere furnishes the necessary conditions for the existence of plants and animals, which in themselves are geologically important.

THE MECHANICAL WORK OF THE ATMOSPHERE

The mechanical work of the atmosphere is accomplished chiefly through its movements. A feeble breeze is competent to move particles of dust, a wind of moderate velocity to shift dry sand, and exceptionally strong winds sometimes to move small pebbles.

The principal movement of the wind is horizontal, but every obstacle against which it blows deflects a part of it, and some of these deflections are upward. Furthermore, there are exceptional winds in which the vertical element predominates. Particles of dust are often caught up by these upward currents, carried to great heights, and transported to great distances.

DUST

Small, solid particles in suspension in the atmosphere are called dust. The transportation of dust by the wind is nearly universal. No house, no room, and scarcely a drawer is so tightly closed but that dust enters. The visible dustiness of the atmosphere in dry regions during wind-storms is familiar proof of the efficiency of the wind as a carrier of dust.

The sources of dust are manifold. Fine particles of earthy material caught up from dry, exposed surfaces usually predominate. But the solid particles of smoke, minute bits of organic material such as pollen, spores, and bacteria, as well as the fine particles of volcanic dust abundant in the vicinity of explosive volcanoes, make up a considerable fraction of the dust. A small amount of dust comes to the atmosphere from extra-terrestrial sources, for when meteorites burn in the atmosphere they are converted into meteoritic dust.

Under special circumstances opportunity is afforded for rough determinations of the distances and heights to which dust is carried by the wind. Snow taken from snow-fields in high mountain regions is found to contain a small amount of earthy matter. Some of its particles are often volcanic dust, even when the place whence the snow is taken is scores or even hundreds of miles from the nearest volcano. In the great eruption of Krakatoa in 1883 large quantities of volcanic dust (pu'verized lava) were projected to great heights. The coarser particles soon settled; but many of the finer ones, caught by the currents of the upper atmosphere, were carried completely around the earth several times. Through all its long journey, the dust was gradually settling from the atmosphere. But for more than a year enough remained in the atmosphere to produce the historic red sunsets. It is probable that dust from this eruption found its way to nearly all parts of the earth.

The source of volcanic dust, which is shot into the atmosphere rather than picked up by it, is to a certain extent determinable at the time of an eruption, while that of the dust picked up by the wind is less easily determined. It would be an exaggeration, perhaps, to say that every square mile of land surface contains particles of dust brought to it by the wind from every other square mile, but such a statement would probably involve much less exaggeration than might at first be supposed. Extensive deposits of wind-blown volcanic dust, locally as much as thirty feet thick, are found in parts of Kansas and Nebraska, hundreds of miles from the nearest known volcanic vent. Beds of volcanic dust, the source of which must be far distant, are found in many parts of the earth.

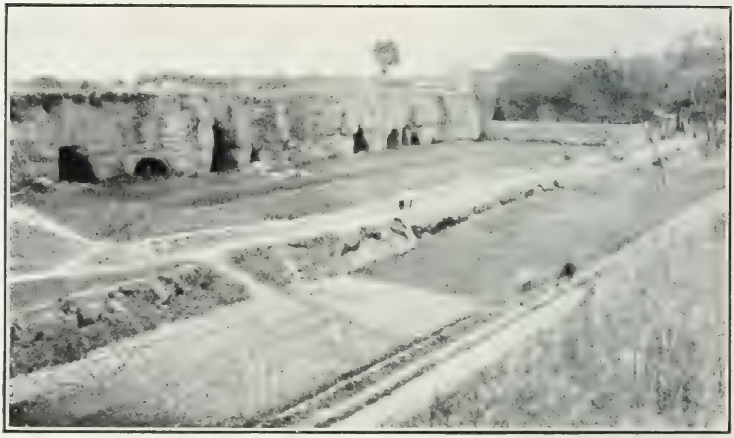

Fig. 10. — Cliffs of loess near Honan Fu, China. Caves made for habitation. (R. T. Chamberlin.)

Loess. In parts of China, in many parts of Europe, and over extensive areas of the Mississippi Basin, there are extensive deposits of earthy material, the particles of which are larger than those of clay but smaller than grains of sand.

This material has been deposited chiefly by the wind and is known as *loess.* The blanket of loess varies from a few inches to a few scores of feet in thickness.

In certain parts of the globe loess is accumulating at the present time. The strong winds from the Gobi Desert blow large quantities of dust onto the neighboring plains of China. During the more severe dust-storms of northwestern China, the sun is often completely hidden by the great quantity of dust in the atmosphere. In many places in China the loess deposits have reached several hundred feet in thickness. From the flood-plains of such rivers as the Missouri, clouds of dust are seen to be swept up and out over the adjacent uplands. Some loess is now being formed in this way, but the great deposits of loess in North America and Europe appear to have been formed largely during the Glacial Period. Where a stream undercuts a bank of loess, the material, due to its compacting and the angularity of its grains, will stand for a long time in vertical cliffs (Figs. 10 and 11).

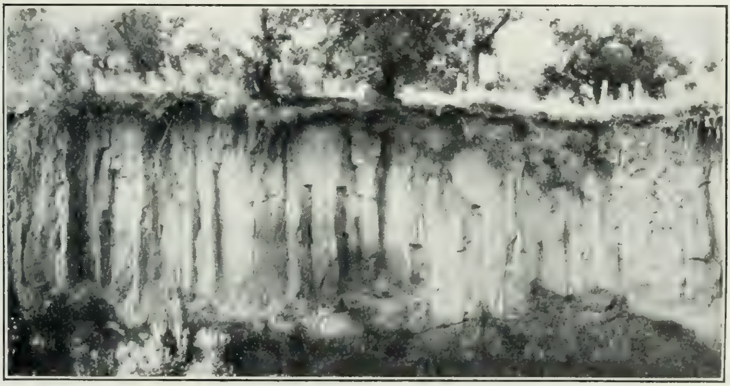

Fig. 11. — Cliff of loess, showing columnar structure. 12 miles east of Leavenworth, Kan. (Albertson.)

Since most dust is carried up in the atmosphere rather than swept along the ground, its movements and its deposition are little affected by obstacles on the surface of the land.

When it settles it is spread more or less uniformly as a blanket.

Gradation. Much dust transported by the wind is blown out over the sea, and settles irretrievably to the bottom. The amount of dust blown into the sea cannot be determined, but it is safe to say that if such determination were possible, the result, stated in terms of weight, would be surprisingly great. In general it may be said that, since land material is being blown into the sea, the tendency is to lower the land and build up the sea bottom.

SAND

When sand of shores, valley bottoms, desert regions, etc., is dry, it is moved by wind of sufficient velocity. The wind does not commonly lift it high above the surface, but rather rolls and bounces the grains along. The movement of the material is therefore interfered with by any surface obstacles which check the velocity of the wind. A shrub, a tree, a fence, a building, or even a stone may cause the lodgment of sand.

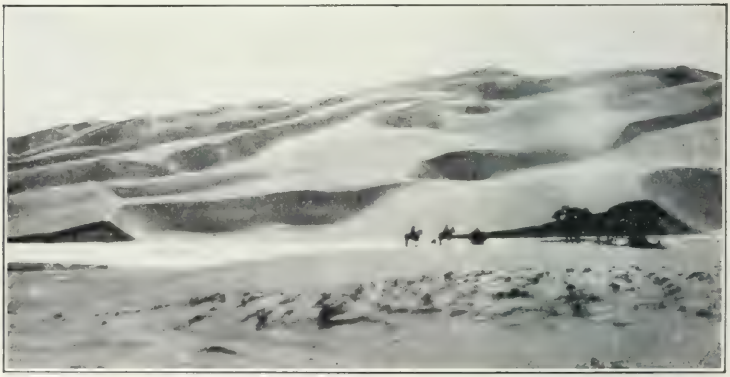

Fig. 12. — The Great Sand Dunes, San Isabel National Forest, Colo. (Courtesy of U. S. Forest Serv.)

Dunes. It is evident that sand, instead of being deposited like dust in a more or less even covering, accumulates in mounds and ridges, which begin their growth against, or in the lee of, some obstacle on the surface. Such mounds or ridges of wind-blown sand are called *dunes*. Once started, dunes may continue to grow by the lodgment of more sand so that they may reach a height of 200 or 300 feet, though commonly they are only 10 to 20 feet high (Fig. 12). Greatest lodgment of sand commonly takes place on the lee side of dunes. Therefore a gentle windward and a steeper lee slope are produced, for on the lee side the angle of slope is the angle of rest of the sand, i.e., the sand has as steep a slope as is possible to it without falling. This angle varies commonly from 25° to 32° depending upon the character of the sand. But since the wind shifts its direction, the slope and shape of a dune also shift from time to time. Also where winds erode (scour) more than they deposit, other profiles are developed and may be very irregular if the dunes are partially covered with vegetation (Fig. 13). One of the results of shifting position and shape of a dune with changing wind conditions, is variation in the direction and slant of the sand layers, which is known as cross-bedding (Fig. 14).

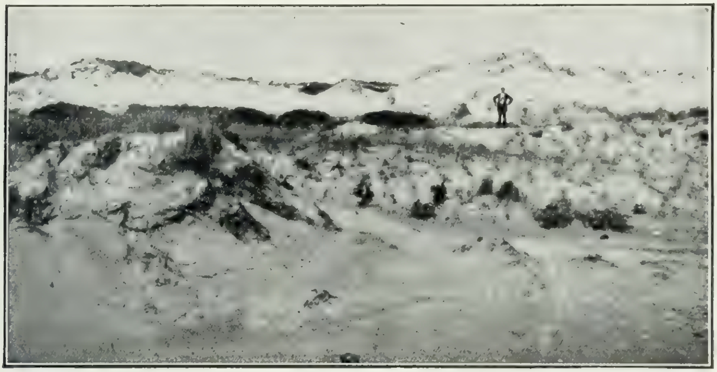

Fig. 13. — Dunes at Longport, coast of New Jersey, showing the irregular forms developed by winds which erode.

The Topographic Map. Since dunes as well as other topographic features are conveniently represented on contour maps, and since such maps will be used frequently in the following pages, a general explanation of them is here introduced.

"The features represented on the topographic maps comprise three general classes — (1) inequalities of surface, such as plains, plateaus, valleys, hills, and mountains, which collectively make up the *relief* of the area; (2) bodies of water, such as streams, lakes, swamps, tidal flats, and the sea, which collectively make up the *drainage;* (3) such works of man as roads, railroads, buildings, villages, and cities, which collectively are known as *culture*.

Relief. All altitudes are measured from mean sea-level. The heights of many points have been accurately determined, and those of some are given on the map in figures. It is desirable, however, to show the altitude of all parts of the area mapped, the form of the surface, and the grade of all slopes. This is done by contour lines, printed in brown, each representing a certain height above sea-level. A contour on the ground passes through points that have the same altitude. One who follows a contour will go neither uphill nor downhill but on a level. The manner in which contour lines express altitude, form, and slope is shown in Figure 15.

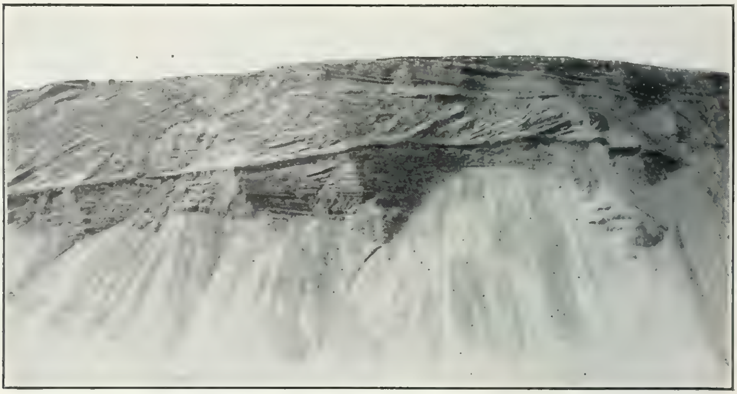

Fig. 14. — Cross-bedding in an Indiana dune. The steam shovel has exposed the irregular stratification.

The view represents a river valley between two hills. In the foreground is the sea, with a bay that is partly inclosed by a hooked sand-bar. On each side of the valley is a terrace. The terrace on the right merges into a gentle upward slope; that on the left merges into a steep slope that passes upward to a cliff, or scarp, which contrasts with the gradual slope back from its crest. In the map each of these features is indicated, directly beneath its position in the view, by contour lines. This map does not include the distant part of the view.

As contours are continuous horizontal lines they wind smoothly about smooth surfaces, recede into ravines, and project around spurs or prominences. The relations of contour curves and angles to the form of the land can be seen from the map and sketch. The contour lines show not only the shape of the hills and valleys but their altitude, as well as the steepness or grade of all slopes.

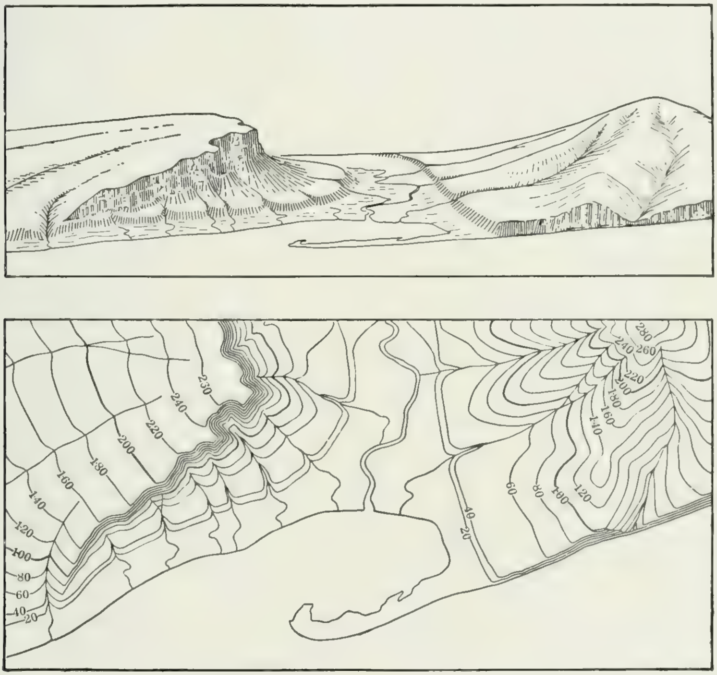

Fig. 15. — Sketch and map of the same area, to illustrate the representation of topography by means of contour lines. (U. S. Geol. Surv.)

The vertical distance represented by the space between two successive contour lines — the contour interval — is the same, whether the contours lie along a cliff or on a gentle slope; but to reach a given height on a gentle slope one must go farther than on a steep slope, and therefore contours are far apart on gentle slopes and near together on steep slopes.

The contour interval is generally uniform throughout a single map. The relief of a flat or gently undulating country can be adequately represented only by the use of a small contour interval; that of a steep or mountainous country can generally be adequately represented on the same scale by the use of a larger interval. The smallest interval commonly used on the atlas sheets of the Geological Survey is 5 feet, which is used for regions like the Mississippi Delta and the Dismal Swamp. An interval of 1 foot has been used on some large-scale maps of very flat areas. On maps of more rugged country contour intervals of 10, 20, 25, 50, and 100 feet are used, and on maps of great mountain masses like those in Colorado the interval may be 250 feet.

In Figure 15 the contour interval is 20 feet, and the contour lines therefore represent contours at 20, 40, 60, and 80 feet, and so on, above mean sea-level. Along the contour at 200 feet lie all points that are 200 feet above the sea — that is, this contour would be the shore line if the sea were to rise 200 feet; along the contour at 100 feet are all points that are 100 feet above the sea; and so on. In the space between any two contours are all points whose altitudes are above the lower and below the higher contour. Thus the contour at 40 feet falls just below the edge of the terrace, and that at 60 feet lies above the terrace; therefore all points on the terrace are shown to be more than 40 but less than 60 feet above the sea. In this illustration all the contour lines are numbered, but on most of the Geological Survey's maps only certain contour lines — say every fifth one, which is made slightly heavier — are numbered, for the heights shown by the others may be learned by counting up or down from these. More exact altitudes for many points are given in bulletins published by the Geological Survey.

Drainage. Watercourses are indicated by blue lines. The line for a perennial stream is unbroken; that for an intermittent stream is dotted; and that for a stream which sinks and reappears is broken. Lakes and other bodies of water and the several types of marshy areas are also shown in blue.

Culture. Symbols for the cultural features and for public-land lines and other boundary lines, as well as all the lettering and the map projection, are printed in black." (U. S. Geol. Survey Folios.)

Topography of Dune Areas. From what has been said, it is clear that the topography of dune regions, though always distinctive, may vary widely. Where the dunes take the form of ridges (Fig. 1, Pl. 1), they are often of essentially uniform height and width for considerable distances. If there are parallel ridges they are often separated by trough-like depressions. Where dunes assume the form of hillocks (Figs. 2 and 3, Pl. 1), rather than ridges, the topography is even more distinctive. In some regions depressions (basins) are associated with the dunes.

Explanation of Plate 1. In Figure 1, Plate 1 (Five Mile Beach, 8 miles northeast of Cape May, N. J.) the contour interval is 10 feet. There is here but one contour line (the 10-foot contour), though this appears in several places. Since this line connects places 10 feet above sea-level, all

PLATE I

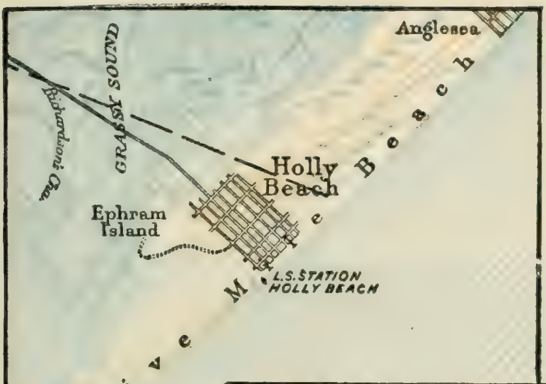
Anglesea
GRASSY SOUND
Richardson's Cha.
Holly Beach
Ephram Island
L.S. STATION HOLLY BEACH
Five Mile Beach

FIG. 1.—Dunes on coast of New Jersey. Scale, about 1 mile per inch. (Cape May Sheet, U. S. Geol. Surv.)

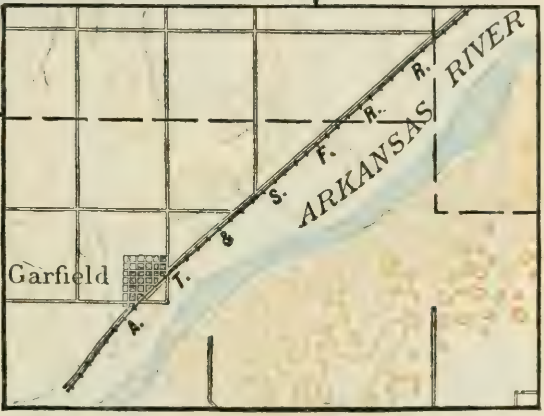
ARKANSAS RIVER
A. T. & S. F. R. R.
Garfield

FIG. 2.—Dunes along Arkansas River in Kansas. Scale, about 2 miles per inch. (Larned Sheet, U. S. Geol. Surv.)

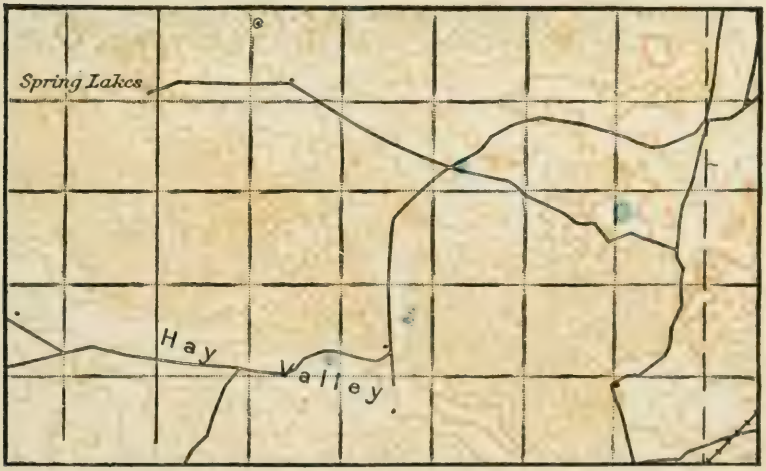
Spring Lakes
Hay Valley

FIG. 3.—Dunes in plains of Nebraska. Scale, about 2 miles per inch. (Camp Clarke Sheet, U. S. Geol. Surv.)

places between it and the sea (or marsh) are less than 10 feet above the water, while all places within the lines have an elevation of more than 10 feet. None of them reaches an elevation of 20 feet, since a 20-foot contour line does not appear. It will be seen that some of the elevations in Figure 1 are elongate, while others have the forms of mounds. (From Cape May, N. J., Sheet, U. S. Geological Survey.)

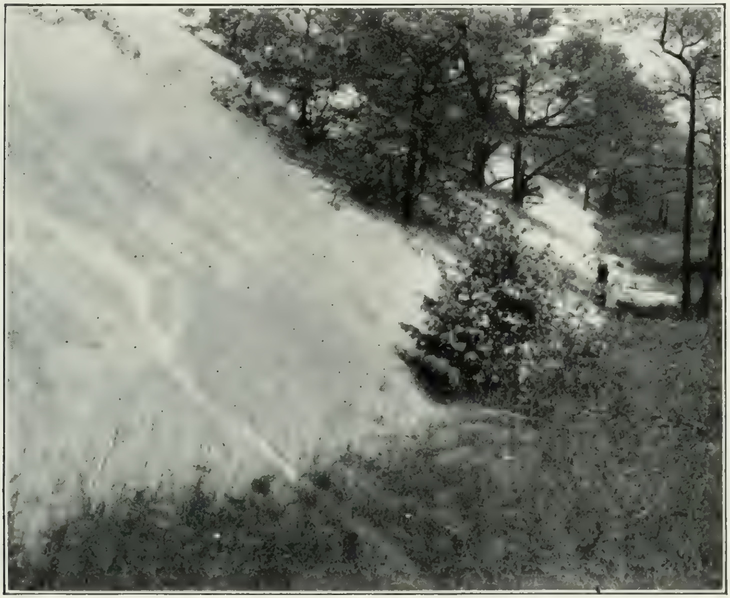

Fig. 16. — Showing how dunes migrate. Indiana dunes.
(Courtesy of U. S. Forestry Serv.)

Figure 2 shows dune topography along the Arkansas River in Kansas (Larned Sheet), and Figure 3, dune topography in Nebraska (Camp Clarke Sheet), not in immediate association with a valley or shore.

In Figure 2 the contour interval is 20 feet. All the small hillocks southeast of the river are dunes. Some of them are represented by one contour, some by two. In Figure 3, where the contour interval is also 20 feet, there are, besides the numerous hillocks, several depressions (basins). These are represented by hachures inside the contour lines. In some cases there are intermittent lakes (blue) in the depressions. There are two depression contours (4280 and 4260) within the contour of 4300, near Spring Lake. The bottom of the depression is therefore lower than 4260, but not so low as 4240.

Migration of Dunes. By the continual transfer of sand from its windward to its leeward side, a dune may be moved from one place to another, though continuing to be made up, in large part, of the same sand. In the migration, dunes sometimes invade fertile lands, causing so great loss that means are devised for stopping them. The simplest method is to help vegetation to get a foothold on the windward side in order to hold down the sand. As a dune ridge along a coast travels inland, another may be formed behind it, and successions of dune ridges are thus formed. When dunes migrate into a timbered region, they bury and kill the trees. In one instance on the coast of Prussia a tall pine forest, covering hundreds of acres, was destroyed during the brief period between 1804 and 1827. In some places in New Jersey orchards have been so far buried within the life-time of their owners that only the tops of the highest trees are exposed. Trees and other objects once buried may be uncovered (resurrected) by further migration of the sand (Fig. 17).

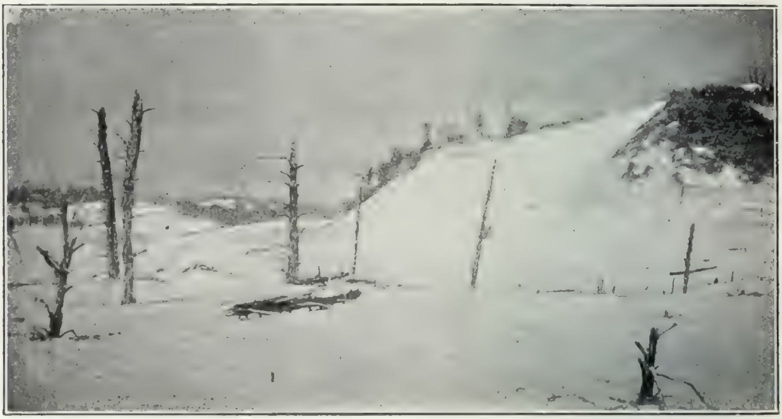

Fig. 17. — A resurrected forest. After burying and killing the forest, the sand was blown away, exposing the dead trees. (Myers.)

Distribution of Dunes. Dunes are likely to be developed wherever dry sand is exposed to the wind. They are especially characteristic of the dry, sandy shores of lakes and seas, of sandy valleys, and of arid, sandy plains (Fig. 18).

Along coasts, dunes are likely to be developed only where the prevailing winds are on-shore. Thus about Lake Michigan, where the prevailing winds are from the west, dunes are abundant and large on the east shore, and but few and small on the west shore. Along valleys, dunes are usually on the side toward which the prevailing winds blow. Dunes may be formed in the valley bottoms, but the sand is often blown up out of the valley, and lodged on the bluffs above.

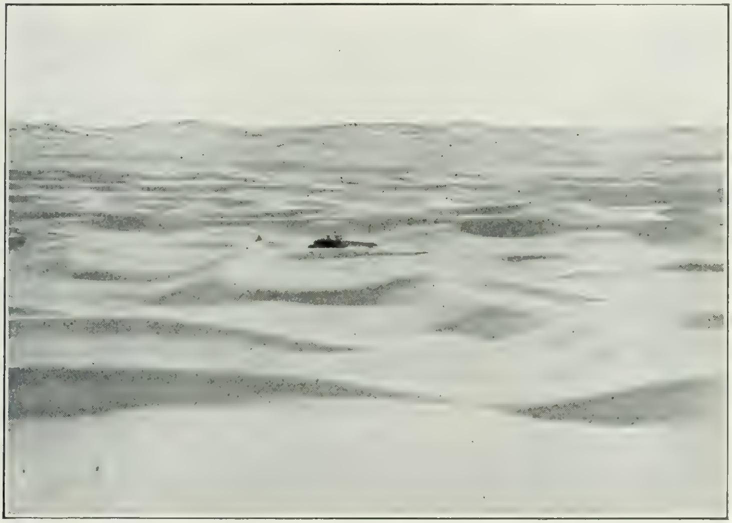

Fig. 18. — The Sahara. (World Wide Photos.)

Eolian sand is not invariably built into dunes. It is in some places spread fairly evenly over the surface. Eolian sand is probably much more widespread than are dunes.

Wind Ripples. A surface of dry sand over which a moderately strong wind has blown for a few hours is likely to be marked with ripples (Fig. 19). While the ripples are, as a rule, but a fraction of an inch high, they throw much light on the migration of great dune ridges. If the ripples be watched closely while wind is blowing, they are seen to shift their position gradually. Sand is blown up the gentler windward slope to the crest of the ripple, and falls down on the other side. Wear on the windward side, about equal to deposition on the leeward side, results in the orderly progression of the ripples in the direction in which the wind is blowing, just as in the case of dune ridges.

Abrasion by the Wind. Sand and dust blown by the wind slowly wear even hard rock. Rock worn in this way acquires a pitted, frosted, or polished surface peculiar to the agency accomplishing the work. If different parts of the rock surface against which the sand is driven differ in hardness, the blown sand digs out the softer parts, leaving the harder portions projecting. The sculpturing thus effected on projecting masses of rock, in many places picturesque and striking (Figs. 20 and 21), is most common in arid regions. So effective an eroding agent is wind-blown sand that in desert regions it is necessary to protect the telephone poles by piling stones to a height of a foot or two about their bases.

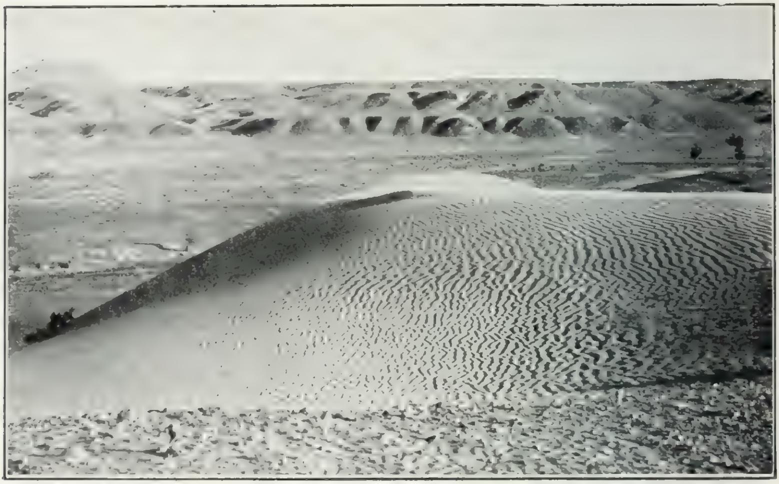

Fig. 19. — A ripple-marked sand dune in a western valley. (U. S. Geol. Surv.)

CHEMICAL WORK OF THE ATMOSPHERE

Of the constituents of the atmosphere, three, oxygen, carbon dioxide, and water, are chemically very active and produce important geologic effects. The activity of oxygen is seen in many familiar phenomena. When a piece of iron or steel, such as a knife blade, is exposed to the air, in the presence of sufficient moisture, the iron rusts and crumbles away. This means that oxygen and water have entered into combination with the iron. The union of oxygen with other substances is *oxidation*, while the union of water with other substances is *hydration*. In the case of iron rusting, a hydrated iron oxide is produced. Other elements besides iron may become oxidized and hydrated. In like fashion the carbon dioxide (CO_2) of the atmosphere unites with many elements to form carbonates. This process is *carbonation*.

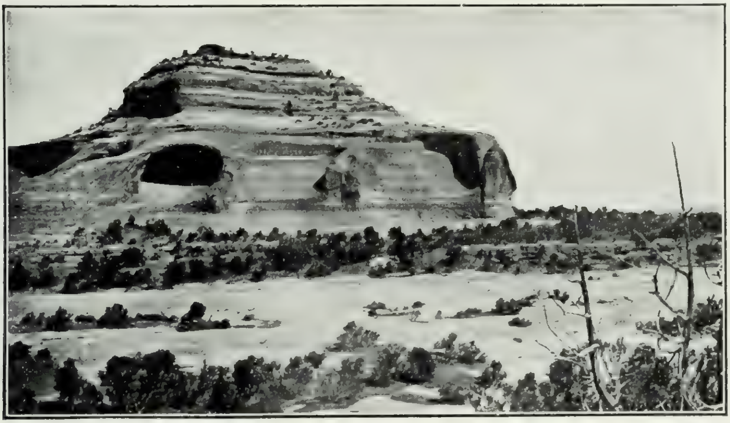

Fig. 20. — Wind erosion. Cave rocks near Sierra La Sal, in Dry Valley, Utah. (Cross, U. S. Geol. Surv.)

The processes of oxidation, hydration, and carbonation are constantly going on where rock is exposed to damp air. Iron, usually in some combination, is a common constituent of many rocks, and much of this iron is subject to the same change as that noted in the knife blade. The oxidation and hydration of this iron causes an increase of volume which leads to the crumbling of the rock. Certain portions of the calcium, magnesium, iron, and certain other constituents of many rocks, are susceptible of carbonation and this process also tends to decompose the rocks. Expansion of the outer part of a rock by these chemical changes, while the subsurface portion remains practically unaffected, frequently produces a "shelling off," or exfoliation (Fig. 22).

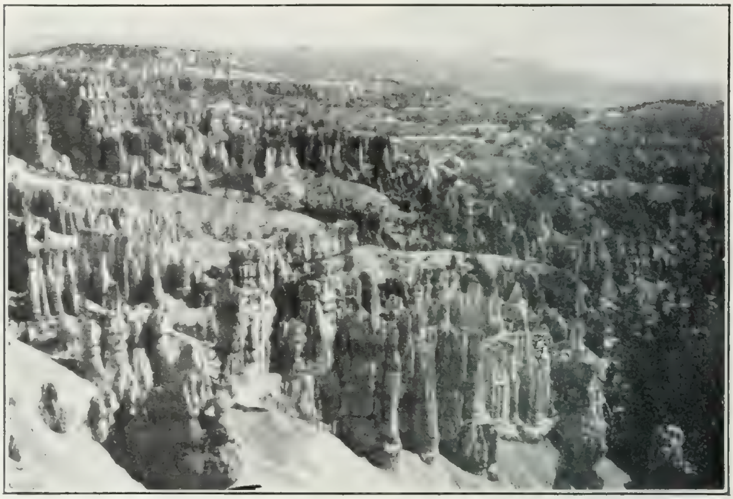

Fig. 21. — Effects of weathering and wind-sculpturing. Bryce Canyon, Utah. (Courtesy of U. S. Forestry Serv.)

Again, some of the compounds, especially the carbonates formed when the rock decays, are soluble and may be dissolved and taken away. This process tends to make the rock less compact and eventually to crumble.

Oxidation, carbonation, and hydration, by changing the chemical nature of the rock, allow some of its material to be carried off in solution, in most cases causing it to fall to pieces. The result is decayed rock, or one variety of mantle rock.

Since chemical decomposition is greatest in warm regions, and the products of decay are least readily removed where there are forests, the products of rock decay are commonly deepest in warm, forested regions. They are very deep, for example, in some parts of Brazil and of Africa.

Weathering. All these changes by which rock is caused to crumble or "decay" are phases of the general process of *weathering*. It is chiefly by this process that most soil, the greatest economic resource of the earth, is produced.

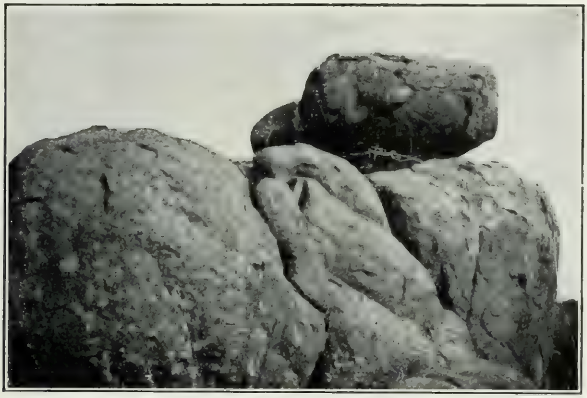

Fig. 22. — Exfoliation of granite. Wichita Mountains, Okla.

THE ATMOSPHERE AS A CONDITIONING AGENT

Though the most obvious mechanical work of the atmosphere is effected by the wind, mechanical results conditioned by the atmosphere, even when the air is still, are also important.

Mechanical Disruption. The chief agent of mechanical disruption is change of temperature. Heating by day and cooling by night produce some such change in rock as that effected by the sudden heating of cold glass or the sudden cooling of hot glass. Since rock is a poor conductor of heat, the outer shell only expands under the influence of the sun's heat. A strain in this way is set up between the outer part and that beneath, which may lead to disruption. This disruption, known as spalling, is most conspicuous on the bare rocks of mountain tops and deserts. If the pores and cracks of the rock are full of water, a fall of temperature below the freezing point is particularly effective. Since water expands about one-tenth its volume on freezing, it acts like a wedge, widening the cracks and disrupting the rock. This *wedge-work of ice* is naturally most important in regions where there are frequent fluctuations of temperature past the freezing point.

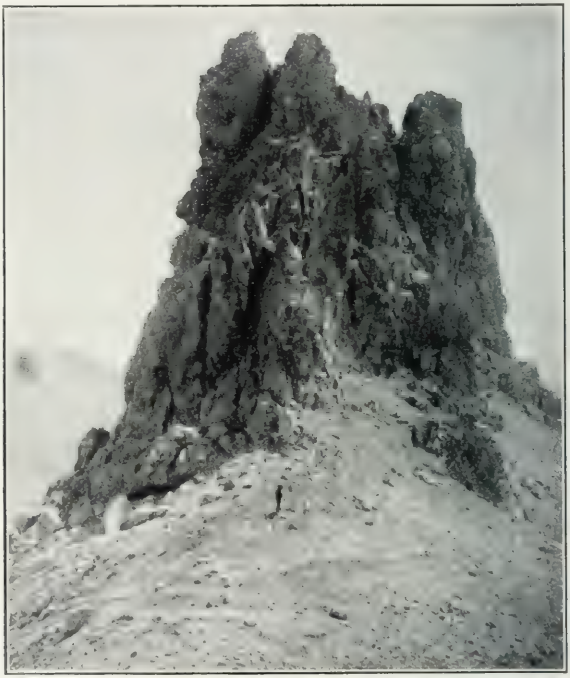

Fig. 23. — A mountain top, showing a common condition of the rock in the mountain peaks.

The importance of these methods of rock breaking is rarely appreciated except by those familiar with high regions and dry regions. Mountain climbers know that most high peaks are shattered, and in cases covered with broken rock to such an extent that it is dangerous for the uninitiated to ascend them. High serrate peaks, especially of crystalline rock, are, as a rule, literally crumbling to pieces (Fig. 23). The piles of *talus* which lie at the bases of steep slopes (Fig. 24) are in places hundreds of feet high. Their materials are often found to be, in large part, the result of the processes here under discussion. While small pieces of rock are more commonly detached in this way, it sometimes happens that large masses, scores and even hundreds of pounds in weight, are dislodged and started on their downward courses.

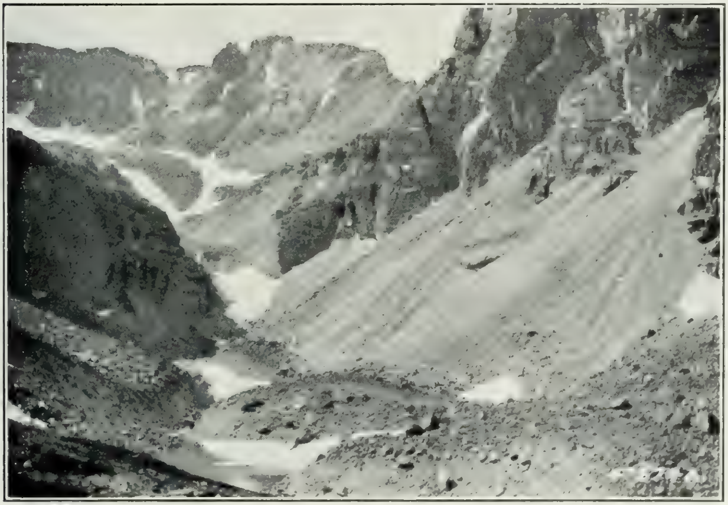

Fig. 24. — Talus, produced by weathering. Glacier Lake and approach to St. Vrain Glacier, Colo. (Courtesy of U. S. Forestry Serv.)

Other examples of weathering under the influence of the atmosphere may be seen. Even in low latitudes and moist climates the effects of temperature changes are evident. For example, thin beds of limestone on the bottoms of quarries are sometimes so expanded by the heat of the sun that they arch up and break.

Instances are known where rocks have been fractured by a stroke of lightning and large masses moved appreciable distances, but in the aggregate the result is inconsequential. Roots of plants not only loosen the soil but wedge rocks apart by growing in cracks. Considerable masses of rock have been pried asunder in this way. When trees are uprooted, soil and rock are displaced. Products of decaying vegetation aid in dissolving and decomposing the rock. Burrowing animals of all kinds loosen the ground.

SUMMARY

On the whole, the tendency of the atmosphere and of all work which is controlled by it is to degrade the land, and to loosen the materials of the surface so that they may be readily moved to lower levels by other agents. The most important phase of the degradational work of the atmosphere is *weathering*, or the preparations of material for removal by other and more powerful agents of degradation. As we shall see later, however, the atmosphere is but one of the agents concerned in weathering.

The wind has doubtless been an important agent in the transportation of dust and sand, wherever and whenever there was dry land, ever since an atmosphere has existed. If it has been as effective, as it is today, through all the millions of years since there have been land and atmosphere, the total amount of work which it has done is stupendous. And there is reason for believing that it has always been effective, since great deposits of wind-blown sand, now cemented into solid rock, have been identified, even in very ancient formations.

WEATHER AND CLIMATE

Weather and climate, in their effects on geologic processes, have played an important rôle in the history of the earth. To understand these effects we must study (1) atmospheric temperature, (2) atmospheric moisture, (3) atmospheric pressure, (4) atmospheric circulation, (5) weather, (6) climate, and (7) climates of geologic time.

1. ATMOSPHERIC TEMPERATURE

Heating. The greater part of the heat which warms the atmosphere comes from the sun. In the transfer of this heat to the atmosphere, three processes are involved. These are *radiation, conduction,* and *convection.*

1. *Radiation.* The process by which heat is transferred through space without the aid of a tangible medium is *radiation.* This heat manifests itself only when absorbed by matter. Sunshine in winter feels warm to the hand even though it may have passed through the cold air and a cold window-pane, because the hand absorbs the radiant heat while the air and the window-pane transmit most of it. Likewise, an object in front of a fire becomes warm on the near side even though the surrounding air be chilly. The land is warmed by absorbing heat radiated from the sun by day, and cools by radiating away this heat by night.

2. *Conduction.* If one end of a piece of iron is heated, the heat is conducted through the iron till the other end becomes hot. It is by this process of conduction that the bottom of the air is warmed where it is in contact with the warm land.

3. *Convection.* In heating gases and liquids *convection* currents are produced. Suppose a vessel of water (Fig. 25) were heated in the middle of the bottom. (*a*) The water, directly over the source of heat, is heated by conduction and so expanded. (*b*) This raises the surface of the water. (*c*) From this raised portion in the center, water flows outward to level the surface. (*d*) There is now more water on the sides than there is in the center. (*e*) There is therefore more pressure at the bottom near the sides than near the middle, which causes the water to crowd in and push up the middle column. Thus *convection circulation* is established.

The surface of the land, warmed by the sun, heats the air above it and produces convection. It is not necessary for the expanded air actually to rise to the upper surface of the atmosphere, for as it expands upward it compresses the air above the lower heated part (Fig. 26). From this compressed area air flows outward. The heavier columns of air in the adjoining regions consequently force air back into the lower part of the heated portion and thus establish convection. This, actually taking place in the air, is probably an important cause of atmospheric circulation.

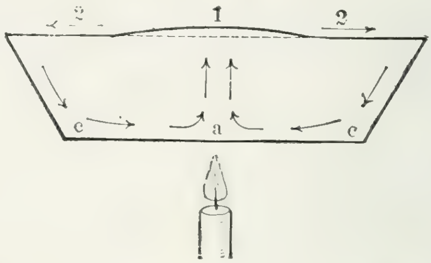

Fig. 25. — Diagram to illustrate convection in a vessel of water heated at one point at the bottom. Surface dome greatly exaggerated.

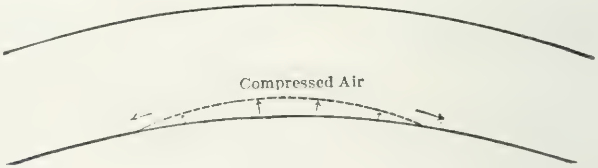

Fig. 26. — Flow of air from above a heated area, which takes place even if the surface of the air were not raised.

Unequal Heating. The total amount of heat received from the sun is fairly constant, but several factors cause this heat to be unequal in its effect.

1. *Day and Night.* Due to the rotation of the earth upon its axis, half of the globe is being warmed by the sun while the other half is being cooled by radiation.

2. *Seasons.* Due to the revolution of the earth about the sun and the inclination of its axis ($23\frac{1}{2}^\circ$) to the plane

of its orbit (Fig. 27), the two hemispheres receive different amounts of heat at the time of the solstices. In the winter hemisphere the sun's rays strike the earth at a low angle, the days are short, and the amount of heat (insolation) small compared with the summer, when the sun is high and the days are long.

3. *Land and Water.* Due to the facts (*a*) that land has lower *specific heat*[1] than water, (*b*) that land absorbs more of the sun's rays than water, which reflects a considerable part of them, (*c*) that land is solid while water has currents to disperse the heat, (*d*) that land is cooled less by evaporation than water, and (*e*) that land is impenetrable to the rays of the sun, the surface of the land becomes hotter under insolation than does that of water; and conversely, land cools more readily by night or in winter than does water.

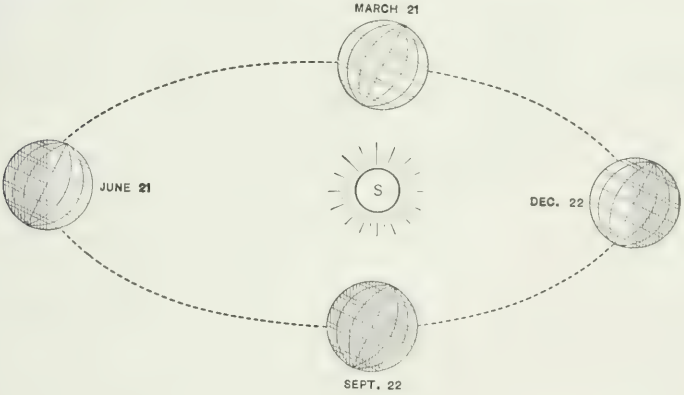

Fig. 27. — Diagram showing the position of the earth and of its insolation at the solstices and equinoxes.

4. *Altitude.* Due to the rarefaction of the air and the consequent diminution of its thermal blanketing effect, high altitudes, on the average, are colder than low ones. The temperature normally decreases about 1° F. in 300 feet of rise. When gas is expanded it is cooled, and when com-

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[1] A given amount of heat will raise the temperature of rock more than that of water.

pressed, it is warmed. So when air rises, expanding with reduced pressure, it cools. Dry air normally cools about 1° F. for every 183 feet it rises. Moisture in the air diminishes this rate of cooling for reasons which will appear later.

Thermal Charts. The distribution of temperature is most easily shown on a *thermal chart or map* on which all points having any given temperature are connected by a line, called an *Isotherm*. There may be annual isotherms, monthly isotherms, etc. A chart or map showing the distribution of the isotherms is an isothermal chart. (Note: Measured temperatures are reduced to sea-level equivalents when recorded on maps.)

Study of the January and July isothermal charts will bring out the general facts that (1) there is a general relationship between isotherms and latitude, (2) there is shift of the isotherms with the seasons, (3) there is notable bending of the isotherms in passing from land to water, and that (4) ocean currents affect the courses of isotherms.

II. ATMOSPHERIC MOISTURE

Water is present in the atmosphere, though in varying amounts, at all times and in all places. It is present in the form of vapor and becomes perceptible only when condensed into dew, fog, clouds, rain, or snow. Its source is evident from the familiar fact of evaporation from bodies of water and from all moist surfaces. But water also evaporates from land which *seems* dry for, below the surface, water is rising and evaporating into the air. Volcanic eruptions also add great quantities of water to the atmosphere.

Mechanics. The present conception is that water is composed of molecules moving in all directions and with varying speeds. If, by a favorable succession of collisions and rebounds, some particular molecule acquires sufficient speed to overcome the attractions of the other molecules in the liquid, it may escape from the liquid, i.e., *evaporate*.

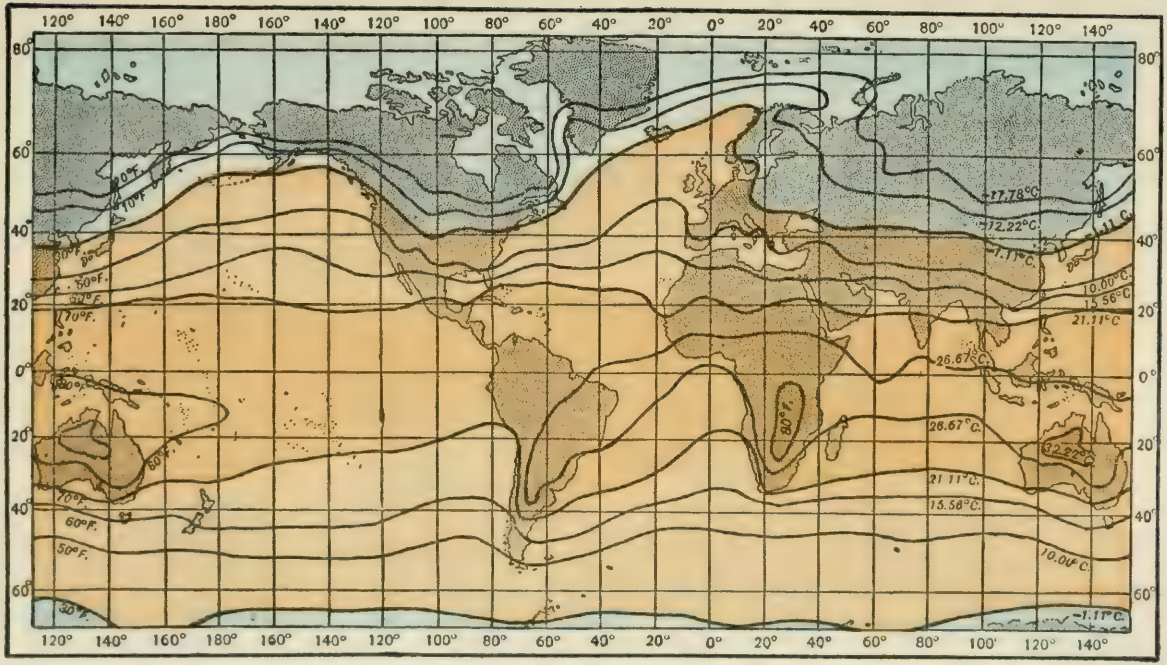
-17.78°C.
-12.22°C.
-1.11°C.
-7.11°C.
10.00°C.
15.56°C.
21.11°C.
26.67°C.
26.67°C.
32.22°C.
21.11°C.
15.56°C.
10.00°C.
-1.11°C.
80°F.
0°F.
10°F.
30°F.
50°F.
60°F.
70°F.
80°F.
80°F.
80°F.
70°F.
60°F.
50°F.
30°F.

FIG. 28.—Isothermal chart for January. (After Buchan.)

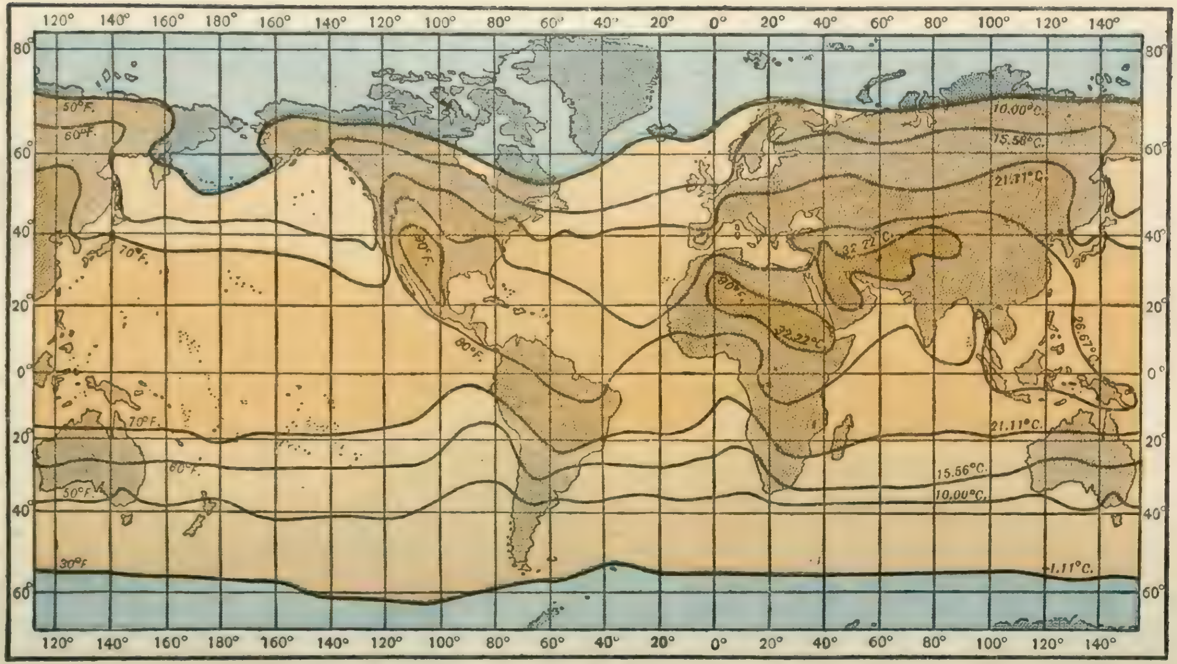
120°
140°
160°
180°
160°
140°
120°
100°
80°
60°
40°
20°
0°
20°
40°
60°
80°
100°
120°
140°
80°
60°
40°
20°
0°
20°
40°
60°
50°F.
60°F.
70°F.
90°F.
80°F.
70°F.
60°F.
50°F.
30°F
10.00°C.
15.56°C.
21.11°C.
32.22°C.
90°F.
32.22°C.
26.67°C.
21.11°C.
15.56°C.
10.00°C.
-1.11°C.

Fig. 29.—Isothermal chart for July. (After Buchan.)

Since the molecules move more rapidly at higher temperatures, it follows that more often they acquire the necessary speed, and more molecules escape from the liquid — in other words the evaporation is more rapid.

As the molecules in the vapor over the liquid move about some may strike into the liquid and remain there. Thus a transfer takes place between liquid and vapor. When equilibrium is established, i.e., as many molecules pass back into the liquid as evaporate from it, the air above the liquid is said to be *saturated*. More accurately, we should say, the space above the liquid is saturated. The actual amount of water vapor in saturated air depends chiefly upon temperature. Warm saturated air contains more water vapor than cooler air.

If warm saturated air be cooled, the speed of the molecules is lessened and *condensation* takes place.

Humidity and Dew-point. The actual amount of water which the air contains is its *absolute humidity*. The percentage of moisture that air contains at any particular temperature, in comparison with the amount that it would have if saturated at that *same* temperature, is the *relative humidity*. The relative humidity of saturated air is 100, while air containing only half as much as it could hold at that temperature, has a relative humidity of 50. The average relative humidity over the land is about 60, while that over the sea is about 85, showing that the atmosphere under average conditions is far from being saturated. Saturation and condensation are produced by special conditions, such as (1) being moved (by wind) to a place where the temperature is lower, i.e., higher altitude or latitude, (2) by cooler air being brought to it, (3) by being cooled by radiation, or (4) by being expanded. When unsaturated air is cooled sufficiently, its relative humidity reaches 100 and the air is saturated. Further cooling causes condensation. The temperature at which condensation begins is the *dew-point*. It is not a fixed point, like the freezing point of fresh water, for if the absolute humidity is high the *dew-point* is high, if low the *dew-point* is low.

Moisture condensing from the atmosphere takes several forms. In summer and fall evenings, objects on the surface of the ground become cooler than the *dew-point* of the surrounding atmosphere so that moisture condenses on them in the form of *dew*. If the dew-point is below the freezing point such condensation produces *frost*. When moisture condenses in the atmosphere to finely divided droplets which do not fall, *clouds* and *fog* result. If the drops are large enough to fall they become rain, while if condensation takes place below the freezing point, *snow* results. It is a commonly observed fact that cooling accompanies evaporation, — in other words, that heat is absorbed in the process. And, conversely, heat is liberated to the surroundings by condensation. The damp equatorial regions are not the hottest places of the earth, and the fruit-grower has less fear of a killing frost on a cloudy night than on a clear one.

III. ATMOSPHERIC PRESSURE

Many observable facts show that the atmosphere exerts pressure. If air is pumped out of a thin-walled receptacle, the sides will collapse. This pressure, which is due simply to the weight of all the air above, averages about 15 pounds per square inch at sea-level. With increase in elevation above sea-level the pressure becomes less and less. Atmospheric pressure is measured with a barometer. This instrument, in its simplest form, consists of a glass tube about 36 inches long closed at one end, filled with mercury, and inverted in a dish of mercury. The mercury in the tube falls till the weight of its column is just supported by the atmospheric pressure on the mercury in the dish. Since at sea-level atmospheric pressure commonly holds the mercury up in the tube about 30 inches (760 mm.), the pressure is said to be 30 inches. If pressure becomes less the mercury in the tube falls below 30 inches and if greater it rises above 30 inches.

Pressure is influenced both by the temperature and the humidity of the atmosphere. Warm air is lighter than cold air, and moist air is lighter than dry air.

Isobaric Maps. Imaginary lines drawn through points on the map having the same pressures are *isobars*. A map showing the distribution of isobars is an *isobaric map* or *chart*. Such charts may be constructed to show average pressure [1] for a year, a month, a day, etc.

The annual isobaric chart (Fig. 30) shows (1) a fairly distinct belt of low pressure in equatorial latitudes, (2) two more or less distinct belts of high pressure areas, one about 30° north latitude and one about 30° south latitude, known, from their location just outside the tropics, as the *extra-tropical high pressure belts*, (3) low pressure in the southern hemisphere from about 40° southward, and (4) low pressure areas over the north Atlantic and north Pacific. As the continents are heated more in summer and cooled more in winter than the oceans, this simple arrangement of pressure is modified notably from season to season.

CIRCULATION OF THE ATMOSPHERE

Wind is air in motion near to, and parallel with, the surface of the earth. If the motion is either upward or downward, or is in the upper air, it is not called wind but an *air current*. Winds are designated by the directions *from which* they blow. Thus a northwest wind is blowing from the northwest toward the southeast. Air is caused to move by differences in pressure. Naturally it moves from places of higher to places of lower pressure, and the greater the difference, the stronger the wind. Winds and all but vertical currents are deflected, by the rotation of the earth, to the right in the northern hemisphere and to the left in the southern hemisphere.

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[1] Observed pressures are reduced to sea-level equivalents.

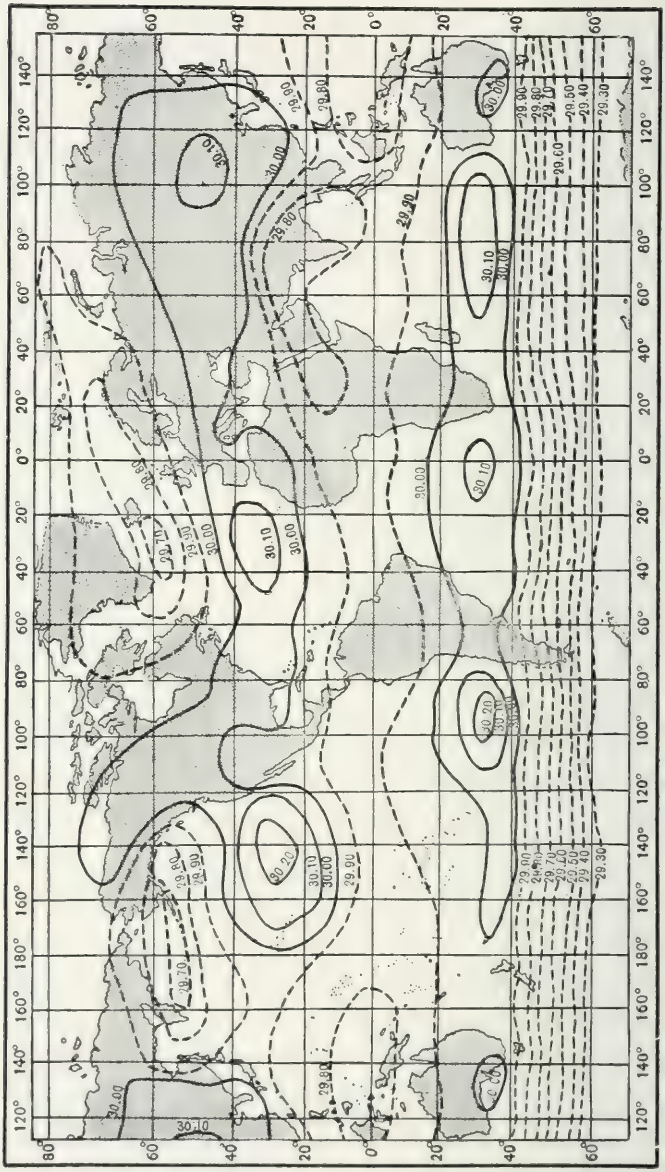
80°
60°
40°
20°
0°
20°
40°
60°
140°
120°
100°
80°
60°
40°
20°
0°
20°
40°
60°
80°
100°
120°
140°
160°
180°
160°
140°
120°
30.10
30.00
29.90
29.80
29.70
29.60
29.50
29.40
29.30
30.20

Fig. 30. — Chart of annual isobars. (After Buchan.)

Suppose that a wind starts straight south from *O* (Fig. 31). At the end of a certain length of time it will have blown a certain distance, say to S^1. But now P^1, in the same longitude as the starting point O^1, is farther east than S^1, due to the increasing circumference of the earth, and hence the greater number of miles traveled in the same time. Hence the wind, by the time it has reached successively the points S^1, S^2, and S^3, even though dragged eastward a little by friction with the surface, is found to have *lagged* behind the longitude of its starting point (P^1, P^2, P^3). This imparts to the wind a component *relative* to the earth's surface, westward in direction. The wind is said to be deflected to the west.

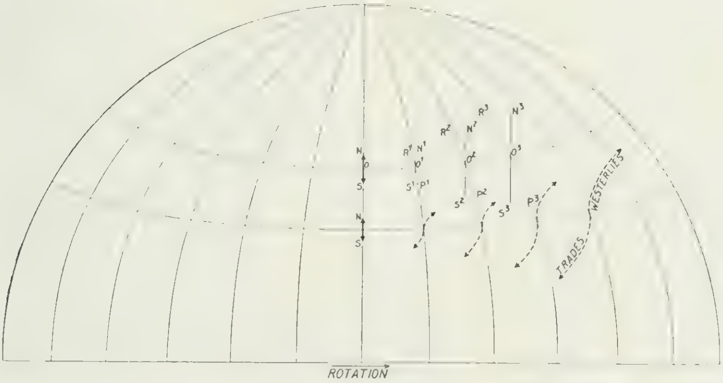

Fig. 31. — Diagram to illustrate the deflection of the winds due to the rotation of the earth.

Similarly a wind starting straight north from *O*, because of its initial easterly rotational component, is found to be successively farther and farther ahead of the positions R^1, R^2, and R^3. This produces an eastward deflection.

Thus the wind in low latitudes blowing toward the equator from the extra-tropical belts is deflected westward to form the *trade winds*, while poleward-blowing winds from the same belts are deflected eastward to form the *prevailing* *westerlies.* But prevailing winds of this kind are interrupted and in cases overcome by winds of other kinds.

(1) Due partly to effects of the upper atmosphere,[1] at present not well understood, and partly to the effects of heating and cooling, areas of low pressure (cyclones) and high pressure (anticyclones) appear at the surface in middle latitudes. These local areas of high and low pressure travel in general from west to east at an average rate of about 700 miles a day, producing the *cyclonic circulation.* Air in the "lows" is rising, being fed by in-blowing winds at the surface, while the air in "highs" is descending to produce out-blowing surface winds.

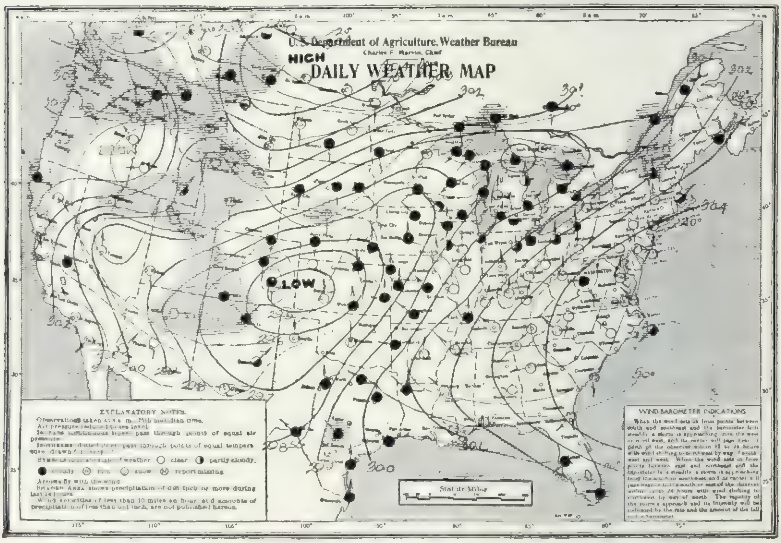
U. S. Department of Agriculture, Weather Bureau
Charles F. Marvin, Chief
HIGH
DAILY WEATHER MAP
LOW
EXPLANATORY NOTES
WIND BAROMETER INDICATIONS
Statute Miles

Fig. 32. — Daily Weather Map of January 9, 1924, showing highs and lows.

(2) Certain well-situated land masses, such as India, are more rapidly heated in summer and more rapidly cooled in winter than the adjacent sea. This results in convection currents which cause winds to blow landward in summer and

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[1] A. E. M. Geddes, *Meteorology.* Blackie and Son, London, 1921, p. 230.

seaward in winter. These winds are the *monsoons*, of preëminent importance to the climate of India.

(3) Daily variations in the heating of land and sea also generate convection currents which cause the wind to blow from the sea to the land, *sea-breeze*, in the daytime, and from the land to the sea, *land-breeze*, at night.

(4) Mountains, because of their exposed position, heat up faster by day and cool off faster by night than the adjoining lowlands. The resulting convection produces winds toward the mountains during the day, *valley breezes*, and away from the mountains during the night, *mountain breezes*.

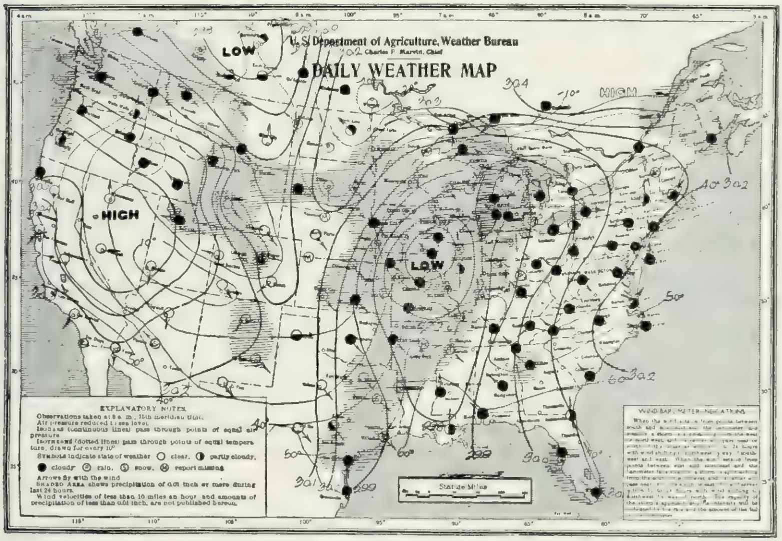

U. S. Department of Agriculture, Weather Bureau
Charles F. Marvin, Chief
DAILY WEATHER MAP
LOW
HIGH
HIGH
LOW
EXPLANATORY NOTES.
Observations taken at 8 a. m., 75th meridian time.
Air pressure reduced to sea level.
Isobars (continuous lines) pass through points of equal air pressure.
Isotherms (dotted lines) pass through points of equal temperature, drawn for every 10°.
Symbols indicate state of weather: ○ clear, ◑ partly cloudy, ● cloudy, Ⓡ rain, Ⓢ snow, Ⓜ report missing.
Arrows fly with the wind.
Shaded Area shows precipitation of 0.01 inch or more during last 24 hours.
Wind velocities of less than 10 miles an hour, and amounts of precipitation of less than 0.01 inch, are not published hereon.
WIND-BAROMETER INDICATIONS
Statute Miles
30.4
30.2
30.0
29.9
40°
50°
60°
70°
-10°
115°
110°
105°
100°
95°
90°
85°
80°

Fig. 33. — January 10, 1924. Storm in Mississippi Valley.

WEATHER

The elements of weather, (1) pressure, (2) winds, (3) humidity, i.e., cloudiness, rainfall, and snowfall, and (4) temperature, are recorded on weather maps.

It is seen that "lows" and "highs" not only control the winds, but affect the temperature, and the former are accompanied commonly by precipitation, while the latter bring fair weather. These highs and lows travel across the continents from west to east at about 700 miles a day (compare Figs. 32, 33 and 34). Therefore, where enough competent observers are at work, weather can be predicted with considerable confidence for as much as 24 hours in advance.

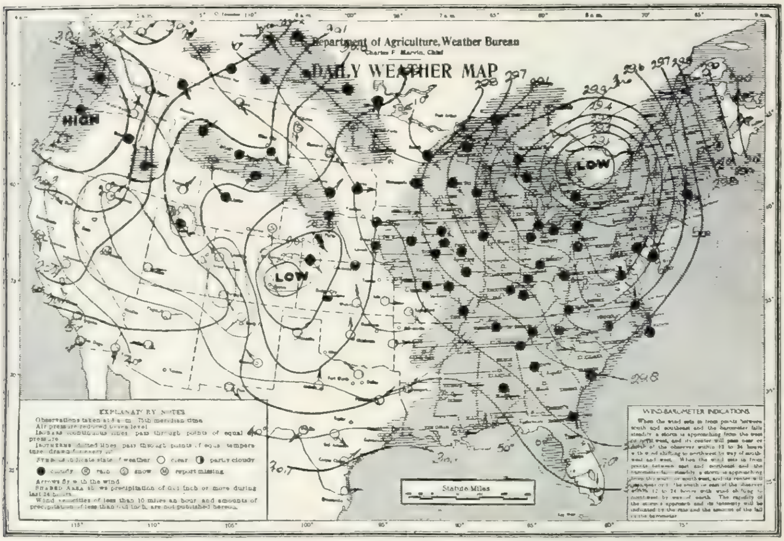

Department of Agriculture, Weather Bureau
Charles F. Marvin, Chief
DAILY WEATHER MAP
HIGH
LOW
LOW
EXPLANATORY NOTES
WIND-BAROMETER INDICATIONS
Statute Miles

Fig. 34. — January 11, 1924. Storm has moved east.

CLIMATIC CONTROLS

The climate of any region is the average of weather conditions over a considerable period of time. In the present state of knowledge it is futile to generalize in genetic climatic classification, but the following explanatory factors may be summarized as the dominating climatic influences of any given region:

(1) *Latitude* is a notable factor in climate.

(2) *Large bodies of water* ameliorate the climate of adjacent lands.

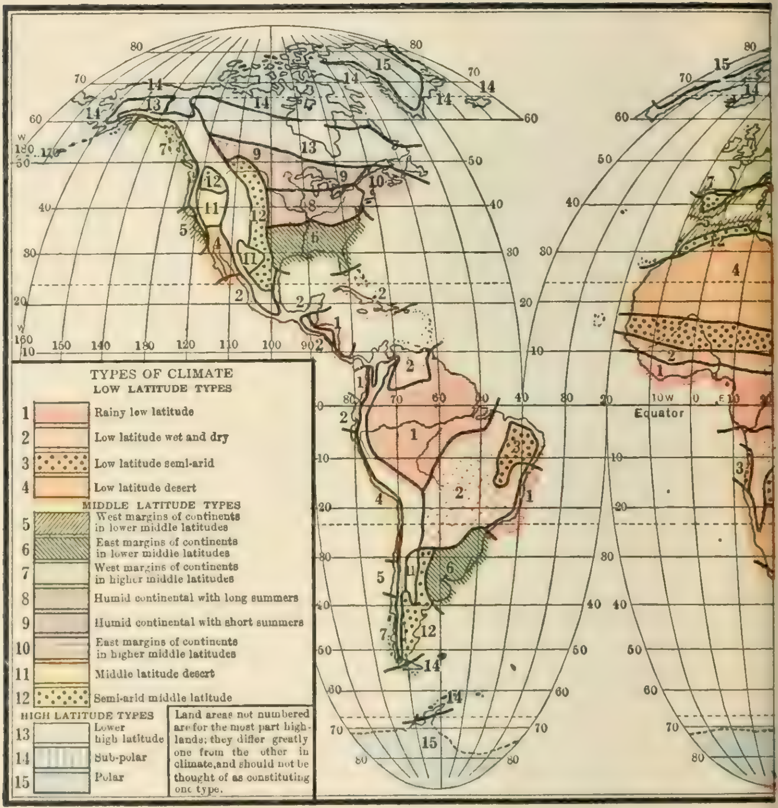
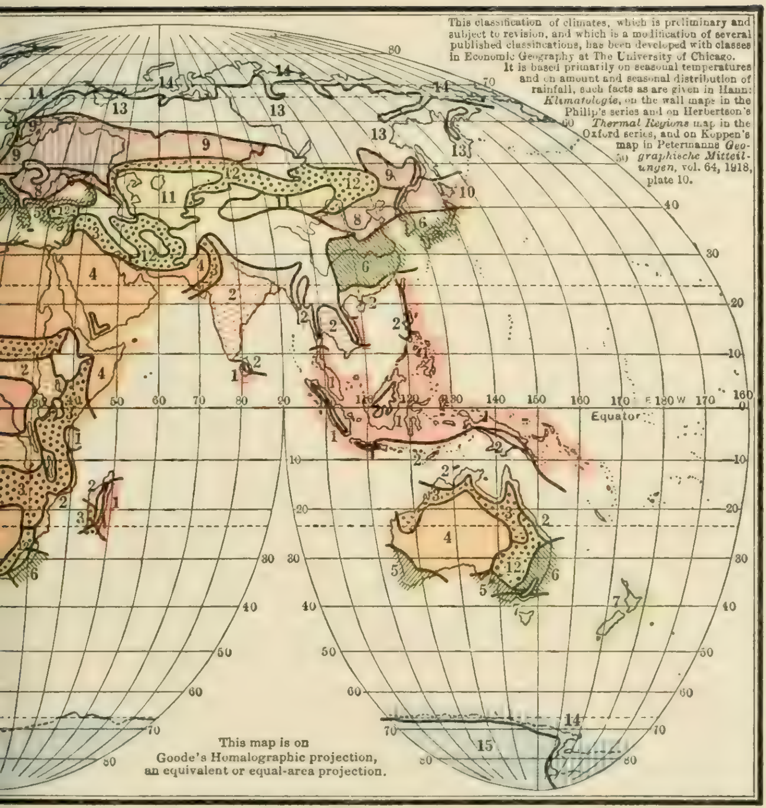

TYPES OF CLIMATE
LOW LATITUDE TYPES
1 Rainy low latitude
2 Low latitude wet and dry
3 Low latitude semi-arid
4 Low latitude desert
MIDDLE LATITUDE TYPES
5 West margins of continents in lower middle latitudes
6 East margins of continents in lower middle latitudes
7 West margins of continents in higher middle latitudes
8 Humid continental with long summers
9 Humid continental with short summers
10 East margins of continents in higher middle latitudes
11 Middle latitude desert
12 Semi-arid middle latitude
HIGH LATITUDE TYPES
13 Lower high latitude
14 Sub-polar
15 Polar
Land areas not numbered are for the most part highlands; they differ greatly one from the other in climate, and should not be thought of as constituting one type.
Equator

Fig. 35. — Climatic Regions


14
13
9
12
11
10
8
6
4
3
2
1
5
7
15
Equator
E 180 W
50
60
70
80
90
110
120
130
140
150
160
170
180
10
20
30
40

This classification of climates, which is preliminary and subject to revision, and which is a modification of several published classifications, has been developed with classes in Economic Geography at The University of Chicago.

It is based primarily on seasonal temperatures and on amount and seasonal distribution of rainfall, such facts as are given in Hann: *Klimatologie*, on the wall maps in the Philip's series and on Herbertson's *Thermal Regions* map in the Oxford series, and on Koppen's map in Petermanns *Geographische Mitteilungen*, vol. 64, 1918, plate 10.

This map is on
Goode's Homalographic projection,
an equivalent or equal-area projection.

From *An Introduction to Economic Geography*, by Wellington D. Jones and Derwent S. Whittlesey

Copyright 1925, by The University of Chicago

f the Continents

(3) In middle latitudes the *cyclonic circulation* is very influential.

(4) *Altitude* influences markedly the climate of a region.

(5) *Mountains* influence climate by causing rainfall on their windward slopes and aridity in their lee.

(6) Other things being equal, *windward coasts* are wetter and have more equable climates than *lee coasts*.

(7) *Monsoons* notably affect the climates of many regions.

(8) It is a growing conviction that *sun-spots* generate stormy conditions and so influence climates.

CLIMATES IN GEOLOGIC TIME

There have been profound changes of climate in the course of the earth's history. The period just antedating the present was one of wide-spread cold climate and glaciation. At several times in geologic history glacial climates are recorded. In fact, during one of the earlier periods of cold, the record shows glaciation in regions which today have tropical and subtropical climates (India, Australia and South Africa), while, on the other hand, at different times, warm climates have existed quite commonly in high latitudes. In fact geologic data indicate that the present climate is much more diversified and in general colder than that during most of geologic history. In addition, great changes of humidity are recorded in the rocks. During one period, for instance, New York and Ohio were arid, while at another time Arizona enjoyed a moist climate and produced a luxuriant vegetation.

In some cases, the cause of climatic variations is local, and due to topographic changes; but in other cases, such as world-wide glaciation, such explanations are not applicable. Various suggestions have been made in the attempt to explain these larger climatic variations. The explanations offered may be grouped as (1) geographic, relating to changes in distribution of land and water, or to topographic changes on the land; (2) astronomic, relating to changes in the earth's orbit, precession of the equinoxes, changing activity of the sun, etc.; (3) atmospheric, relating to the changes in the constitution and circulation of the atmosphere. Though sufficient data are not yet at hand to warrant a final conclusion, accumulating facts point to the third — the atmospheric group of influences — as being the most plausible and the most fruitful line of explanation.

CHAPTER III

GROUND-WATER

The average amount of precipitation on the land is estimated at about 40 inches per year. A part of this water sinks beneath the surface, a part forms pools or lakes on the surface, a part runs off over the surface at once, and a part is evaporated. The proportion of the rainfall in any given place which follows each of these courses depends on several conditions, among which are (1) the topography of the surface, (2) the rate of rainfall (or the rate at which the snow melts), (3) the porosity of the soil or rock, (4) the amount of water which the soil contains when the rain falls or the snow melts, (5) the amount of vegetation on the surface, and (6) the dryness of the atmosphere. The gentler the slopes, the less rapid the rainfall, the more porous the soil, the drier it is, and the heavier the vegetation, the more water will sink beneath the surface.

The water which sinks into the ground becomes *ground-water*, while that which flows off over the surface without sinking in is the *immediate run-off*. Much of the ground-water ultimately reaches the surface again, and some of it joins the immediate run-off in the streams. Whether the water which falls as rain stands on the land as lakes, runs off as rivers, or sinks beneath the surface, it is more or less active, and the results of its activity are seen on every hand.

Many familiar facts demonstrate the presence of abundant water beneath the surface of the land. The thousands of wells made by man, and the many springs which issue from the slopes of mountains, hills, and valleys are sufficient proof both of the wide distribution of ground-water and of its great abundance.

Certain other well-known facts make it clear that ground-water is largely dependent upon rainfall. The level of the water in wells commonly sinks during droughts and rises after rains; and the sinking is greater when the drought is long, and the rise is more notable when the rainfall is heavy. Many springs flow with reduced volume in times of drought and others cease to flow altogether. Furthermore, rain-water is seen to sink beneath the surface wherever the soil is porous. Sinking through the soil to the solid rock, the water finds cracks and pores, and through them it descends to greater depths. Nowhere beneath the mantle rock, when any considerable area is considered, are the rocks so compact and so free from openings, as to prevent the percolation of water through them.

The Ground-water Surface. The Water-table. If wells are sunk in a flat region where the rocks are essentially uniform, the water in them is generally found to stand at a nearly common level. This means that below the common level the pores and cracks are full of water. The highest level below which the rocks are saturated with water is the *ground-water surface* or *water-table*.

In a flat region, of uniform rock structure, this *water-table* is nearly level, though it rises and sinks with rainfall. Where the topography of a region is not flat, the ground-water surface is not level. As a rule it is higher (though farther from the surface) under an elevation than under adjacent lowlands, as illustrated in Figure 36. The reason is readily understood. If a hill of sand is exposed to rainfall, most of the water falling on its porous surface sinks into it. If the precipitation continues long enough, as in a protracted rain, the hill of sand will be filled with water, the water occupying the interstices between the grains. The water in the hill tends to spread, but since the movement involves much friction, the spreading is slow. With the spreading, the surface of the water in the sand sinks, and sinks fastest at the center, where it is highest (same figure). If no more water were added, the surface of the water in the hill would in time sink nearly to the level of the water in the surrounding land. But at every stage preceding the last, the surface of the water would be higher beneath the summit of the hill than elsewhere, though farther from the surface. In regions of even moderate precipitation, the water-table beneath the hills rarely sinks to the level of that in the lowlands about them.

The water-table beneath the lowlands also sinks. Some of the water finds its way into the valleys, some of it sinks to greater depths; some of it evaporates, but since the water-table beneath the elevation sinks more rapidly than that beneath the lowlands, the two approach a common level. Their difference will be least at the end of a drought and greatest after a heavy rain.

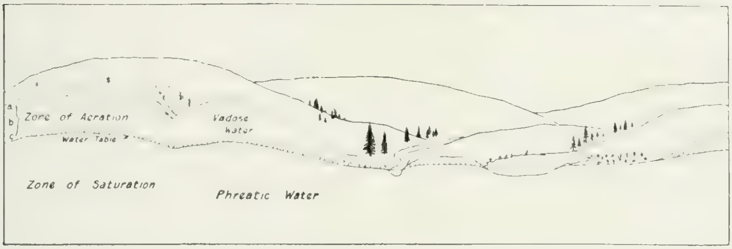


Fig. 36. — Ground-water. Shows relation of water-table to the surface. *a*, belt of soil-water; *b*, intermediate zone; *c*, capillary fringe.

Zone of Aëration. That part of the lithosphere above the zone of saturation is termed the *zone of aëration*,[1] or *vadose zone*. Its interstices are largely filled with air, though suspended water (vadose water) is present in small amount. This zone varies in thickness according to the depth to the water-table. In swamps the zone is absent, while below desert plains it is several hundred feet thick. The average of some 30,000 wells in the United States shows the zone of aëration to be about 37 feet in thickness. This zone may be subdivided into three more or less distinct parts or belts —

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[1] O. E. Meinzer, *U. S. Geol. Survey Water Supply Paper* 489, 1923.

the belt of soil-water, the capillary fringe, and the intermediate belt.

(1) The belt of soil-water includes soil and other materials near enough to the surface to discharge water in appreciable amounts into the atmosphere by evaporation. Convection, capillarity, and plant action assist in bringing water to the atmosphere. Since most plants are dependent for their supply on soil-water, the value of a soil obviously depends to a very large extent on its *specific retention* — capacity for holding water against the pull of gravity. Fine material, like silt or clay, has higher specific retention than sand.

(2) Fine-grained earth is found to be moist for some distance above the water-table. For, through communicating interstices among the grains, capillarity draws moisture up from the zone of saturation. The height to which this water rises varies with the size of the grains, being an inch or so in gravel, and several feet in fine material. This moist zone above the water-table is called the *capillary fringe*.

(3) In places and at times where the soil-water does not extend down to the capillary fringe, there is an *intermediate zone*. This zone below the belt from which water may be withdrawn by evaporation, or plant action, is thick where the water-table is low, and thin or absent where it is high. Water not retained in the soil zone seeps downward into the intermediate zone. Part is there retained, while the rest finds its way to the water-table. That part retained is drawn by capillarity into the smallest openings, where it is held against the pull of gravity. While the average amount of water held in this zone, compared with the rock material, is only a few per cent. by weight, as much as 30 per cent. has been reported in certain fresh clays.

Descent of Ground-water. Water enters the ground through openings between the grains of the mantle rock or through minute pores of the bed-rock. As might be expected, the smaller the cracks or pores, the more difficult is this passage. If the size of an opening be diminished to one-half, the difficulty of passage is more than doubled. Since openings of all kinds in the rocks diminish in size and number, with depth, the amount of ground-water also diminishes and its descent becomes more and more difficult.

Water in deep mines is more abundant in the upper levels than in the lower ones. In many cases, there is apparently little if any circulation of ground-water in the deeper workings. Deep below the surface newly opened workings may be damp, or water may flow from fissures for a short time, but if water from higher levels be excluded, little pumping is required to keep the mine open. Seldom, either, does a well encounter much water below a depth of 2000 feet. While, in a few regions, considerable flows of ground-water are encountered locally, the general fact is evident that the amount of water decreases with increase in depth.

Movement of Ground-water. Ground-water is in more or less continual movement. If a well is pumped dry, it soon fills again by inflow from the sides and bottom. Springs and flowing wells also demonstrate the movement of ground-water. Near the surface the movement is primarily downward if the rock through which it passes is equally permeable in all directions; but so soon as the descending water reaches the water-table its downward flow is checked, and its movement becomes primarily lateral.

Ground-water moves chiefly by slow percolation through the pores of rocks, little of it being organized into definite streams. Small streams are seen in some caves, and subterranean streams issue as springs in some places; but most streams which issue as springs probably have definite channels for short distances only, before they appear at the surface. It is probable that ground-water frequently flows in considerable quantity along somewhat definite zones, without having open channels. This is the case especially along the surface of an impervious layer overlain by a porous one. The "reservoirs" from which artesian wells draw their supply are not usually streams or lakes, but porous beds of rock containing abundant water. As the supply is drawn off at one point, it is renewed by water entering elsewhere. Since the freedom of movement of ground-water is influenced greatly by the porosity of the rock, and since the rock is, on the average, most porous near the surface, the movement of ground-water is, on the whole, greatest near the surface.

Except in very porous soils, the rate of movement ranges from one to eight feet a day. In the Potsdam sandstone of the upper Mississippi Valley region, the rate of movement has been measured to be about one-half mile a year. This means that the water pumped from this formation at Chicago entered this porous bed of rock, where it comes to the surface 150 miles away in Wisconsin, about 300 years ago. This slow movement of water also accounts for the shortage of water in the Dakota sandstone under the Great Plains, the fact being that water is being pumped out faster than it is seeping through the rock from the foothills of the Rocky Mountains. It appears, then, that while the upper part of the ground-water, especially that above the water-table, moves freely, the lower part moves more slowly. It is probable, indeed, that movement in the lower part of the subterranean hydrosphere is extremely slow.

Amount of Ground-water. Since neither the depth to which the ground-water penetrates into the earth, nor the porosity of the rocks below the surface is accurately known, it is possible only to estimate roughly the amount of ground-water. Such estimates, though varying widely, indicate that the ground-water if extracted and spread on the surface would make a layer not more than a few hundred feet deep.

Fate of Ground-water. Most of the water which sinks into the earth reaches the surface again after a longer or shorter journey. Some of it is evaporated from the surface directly; some of it is taken up by plants and is passed by them into the atmosphere; some of it issues in the form of springs; some of it seeps out; some of it is drawn out through wells; and much of the remainder finds its way underground to the sea or to lakes, issuing as springs or seepage beneath them. A small portion of the descending water enters into permanent combination with mineral matter. Many minerals are known to take up water, being changed thereby from an anhydrous to a hydrous condition. It does not necessarily follow, however, that the total supply of water is decreasing. Minerals once hydrated may become dehydrated, the water being set free. Furthermore, considerable quantities of water in the form of vapor issue from volcanoes, and volcanic vents often continue to steam long after active eruption has ceased. It is probable that some, and perhaps much of the water issuing from these vents has not been at the present surface of the earth before, and it is not now possible to affirm that the supply from this source does not offset, or even surpass, the depletion of the hydrosphere resulting from mineral hydration.

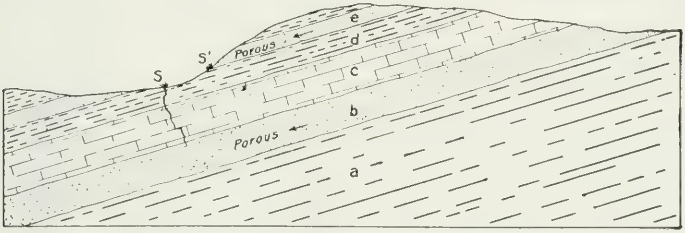

Fig. 37. — Diagram illustrating shallow and deep springs.

SPRINGS

All water issuing from beneath the surface is *seepage.* Water issuing through a natural opening in such quantity as to make a distinct current is a *spring.* Springs occur in many sorts of situations, but they are not located by accident. They occur where there are natural passageways for the ground-water to escape to the surface. Two cases are illustrated by Figure 37. In one, the water descends through the porous bed *e* to the layer *d*, which is relatively impervious. The water flows along this layer until it comes to the surface, and there the water flows out as a spring, *s*′. In the other, the water moves underground through the porous layer *b*, under pressure, until it reaches a crack which leads up to the surface. If the crack is open enough to afford a passageway, the water may follow it up to the surface, as at *s*. A spring will occur in such a situation only where the opening is lower than the water-surface in the rock which furnishes the water. In the figure, the spring at *s* is lower than the water-surface at the intake. This sort of a spring is similar to a flowing well in principle, but in the latter case the opening is made by man.

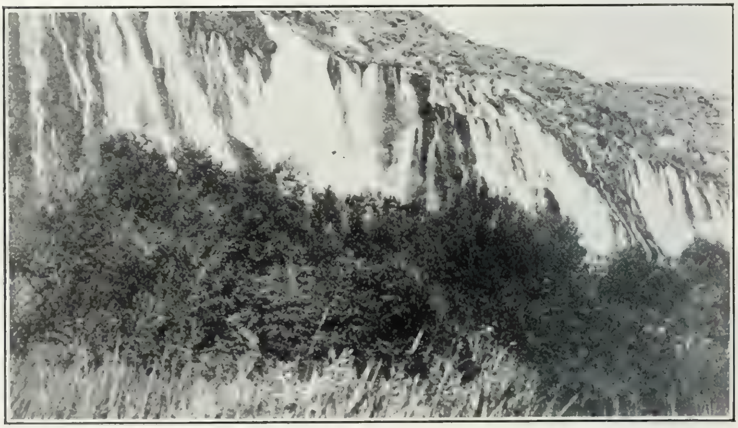

Fig. 38. — Thousand Springs, Idaho, north side of the Snake River. (Russell, U. S. Geol. Surv.)

Temperature. Most springs seem cold in warm weather. There is a popular impression that springs are cooler in summer than in winter, but this is not the case. The impression arises from the fact that the water is much cooler than the air in summer, and so seems cold, while in winter the water is commonly warmer than the air, and so seems less cold than in summer.

Springs which derive their water from deep sources vary little in temperature during the year, while those whose sources are shallow are colder in winter than in summer. The reason is that the warmth of summer and the cold of winter are most extreme at the surface, and become less and less with increasing depth. Below the depth of 50 or 60 feet, in middle latitudes, the temperature does not vary sensibly with the seasons, so that springs which draw their water from greater depths vary little in temperature, while those which draw their supply from lesser depths vary more.

Exceptional springs are warm, and still more exceptional ones are hot. Hot spring water gets its heat by contact with hot rock. In general, the water of warm and hot springs probably rises from considerable depths. In many cases the source of heat probably is igneous rock (lava) which has not yet become cold. It may be lava which was forced upward toward but not to the surface, or it may be the deeper parts of lava which flowed out on the surface.

Classification. Springs are classified in various ways, and the several classifications suggest characteristics worthy of note. They are sometimes classified as *deep* and *shallow*, but the idea involved in this grouping would be better expressed as *strong* and *feeble*. Springs are also classified as *cold* and *thermal*, the latter term meaning simply that the temperature is such as to make the springs seem warm or hot. Between deep springs and shallow ones, and between cold springs and warm ones, respectively, there is no sharp line of demarcation. Again, some springs are continuous in their flow, while others are intermittent. Most intermittent springs flow after periods of precipitation, but dry up during droughts. Springs are also classified as *mineral* and *common*. Mineral springs, in the popular sense of the term, are of two types: (1) those which contain an unusually large amount of mineral matter, and (2) those which contain some unusual mineral. All springs which are not mineral are *common*. This classification is not very significant, for all springs contain more or less mineral matter, and many springs which are common contain more mineral matter than some springs that are called mineral. Mineral springs are themselves classified according to the kind and amount of mineral matter they contain. Thus the *saline* springs contain salt; *sulphur* springs contain compounds (especially gaseous) of sulphur; *chalybeate* springs contain iron compounds, especially the sulphate; *calcareous* springs contain abundant lime carbonate, etc. These various mineral substances have been extracted from the rocks. The salt of saline springs is usually extracted from beds of salt beneath the surface. Lime carbonate, one of the commonest substances in solution in ground-water, is dissolved from limestone or derived by chemical change from rocks containing some calcium compound. The lime feldspars, by carbonation, give rise to lime carbonate. Chalybeate waters often arise from the oxidation of iron sulphide (a mineral which is common in many sedimentary rocks) and the solution of the resulting sulphate. *Medicinal* springs are those which contain some substance or substances which have, or are supposed to have, curative properties.

Mineral Matter in Solution. The number and variety of mineral substances in spring-water are very great, and the amount of solid matter in solution varies widely. Some of the hot springs of Yellowstone Park contain nearly three grams (2.87) of mineral matter per kilogram. The springs of Leuk (Switzerland) bring to the surface annually more than 2000 tons of calcium sulphate (gypsum) in solution, and in the same time the springs of Bath (England) bring up an amount of mineral matter in solution sufficient to make a column 9 feet in diameter, and 140 feet high.

Geysers. Geysers are *intermittently eruptive hot springs*. They occur in the volcanic regions of Iceland and New Zealand, but are best known in Yellowstone National Park.

The important features of a geyser are: (1) a tube or crack, probably crooked or partly obstructed, leading down to (2) hot rock which is probably uncooled lava; (3) a shallow basin about the opening, in some cases in the top of a mound composed of mineral matter deposited from the hot water; and (4) a supply of ground-water to keep filling the tube after it is emptied by the eruptions.

The principles of geyser eruption may be explained as follows: Suppose a geyser-tube full of water to be heated in its lower portion. As the water below is heated convection tends to distribute the heat throughout the column of water above. If the tube were short and convection free, the result would be a boiling spring; but if the tube is long and convection impeded, the water at some level below the surface may be brought to the boiling point before that at the top. Under these circumstances, if even a little water in the lower part of the tube be converted into steam, the steam will so raise the column of water above that it will overflow at the surface. The overflow will relieve the pressure on the heated water. If, before the overflow, there was any considerable volume of water essentially ready to boil, the relief of pressure will allow it to be converted into steam suddenly, and the great expansion involved will cause the eruption of all the water above it.

It is clear that everything which impedes convection in the geyser-tube will hasten the period of eruption, since impeded circulation will have the effect of holding hot water down, and so of bringing the water at some level below the top more quickly to the boiling point. It follows that anything which chokes the tube or impedes the circulation of the water hastens an eruption.

Geysers often build up crater-like basins or cones about their vents, the cone being of material deposited from solution. In Yellowstone Park, the precipitation of matter from solution (chiefly silica) is partly due to cooling, and partly to the algae which abound in water only slightly below the boiling point, and the brilliant colors of some of the deposits about the springs are attributable to these plants.

When the water from any geyser or hot spring ceases to flow, the plants die and the colors disappear.

The heating of geyser-water must cool the lava or other source of heat below. As this takes place, the time between eruptions becomes longer and longer. In the course of time, therefore, the geyser must cease to be eruptive, and when this change is brought about the geyser becomes a hot spring. Within historic time several geysers have ceased to erupt and new ones have been developed. In the Yellowstone Park, where there are said to be something like 3000 vents of all sorts, hot springs which are not eruptive greatly outnumber the geysers. From many of the vents but little steam issues and from some, little else.

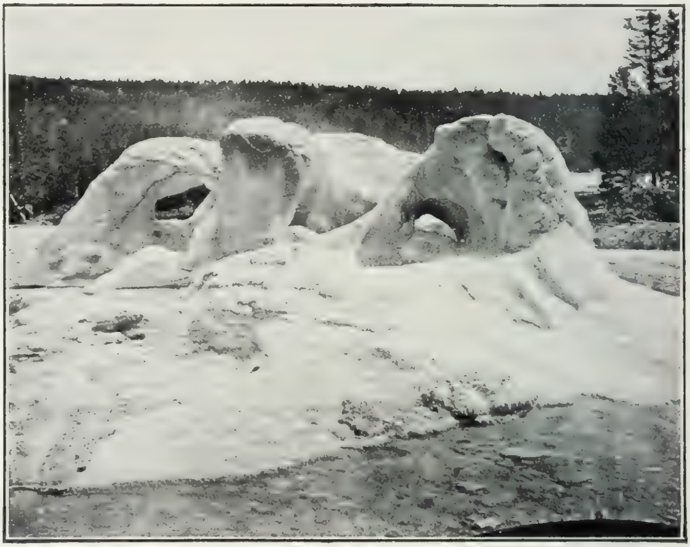

Fig. 39. — Cone (or crater) of Grotto Geyser, Yellowstone Park. (Detroit Photo. Co.)

A few geysers have somewhat definite periods of eruption. Of such, "Old Faithful" is the type; but even this geyser, which formerly erupted at regular intervals of about an hour, is losing the reputation on which its name is based.

Not only is its period of eruption lengthening, but it is becoming irregular, and the irregularity appears to be increasing. In the short time during which this geyser has been under observation, its period has changed from a regular one of 60 minutes or a little less, to an irregular one of 60 to 90 minutes. In the case of some geysers, years elapse between eruptions, and in some, the date of the last eruption is so remote that it is uncertain whether the vent should be looked upon as a geyser or merely a hot spring.

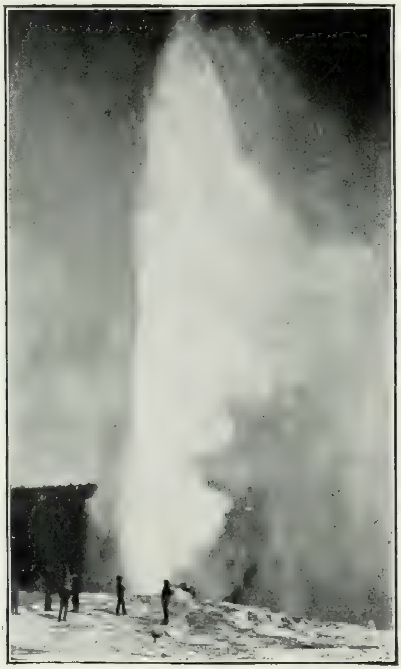

Fig. 40. — Old Faithful in eruption.

In Yellowstone Park, the geysers are mainly in the bottoms of valleys, but the deposits characteristic of geysers are found in several places well above the present bottoms. These deposits record the fact that in earlier times the geysers were at higher levels than now. It is probable that they have been, at all stages in their history, near the bottoms of valleys, and that, as the valleys have deepened, the ground-water has found lower and lower points of issue. In this respect, the geysers have probably had the same history as other springs.

Unless new intrusions of lava occur, or unless heat is otherwise renewed at the proper points, it is probable that all existing geysers will become extinct within a time which is geologically short. New geyser regions may, however, develop as old ones disappear.

Fig. 41. — Norris Geyser Basin. (MacClintock.)

Artesian Wells. A well in which the water rises above the water-table is an *artesian well.* If the water rises so as to flow at the surface a flowing well is formed. Pressure, *artesian head*, to cause this rise is produced by the appropriate relations of rock structure, water supply, and elevation. Generally speaking, a flowing well is possible in any place underlain by any considerable bed of porous rock, if this rock outcrops at a sufficiently higher level in a region of adequate rainfall and is covered by a layer or bed of impervious or relatively impervious rock. This statement involves four conditions, all of which are illustrated by Figure 42, where *a* is the *aquifer* (water transporter). It is not necessary that the beds of rock form a structural basin, nor is it usually necessary to take account of the character of the rock beneath the porous bed which contains the water.

The bed of porous rock is the "reservoir" (aquifer) of the flowing well. Formations of sand or sandstone, and of gravel or conglomerate, most commonly serve as the aquifers. In order that they may contain abundant water they must have considerable thickness, and their outcroping edges must be so situated that the water may enter freely and be replenished, chiefly by rain, as the water flows out at the well.

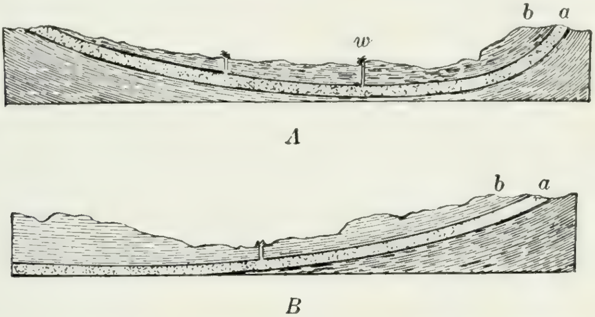

Fig. 42. — Diagrams illustrating conditions favorable for artesian wells. In *A*, the porous bed *a* is in the form of a basin; in *B*, it merely dips.

A relatively impervious layer of rock above the aquifer (*b*, Fig. 42) is most important; otherwise the water in the aquifer will leak out, and there will be little or no "head" at the well-site. Thus if the rock overlaying stratum *a* (Fig. 42) were badly broken, the fractures extending up to the surface, the conditions would be unfavorable for flowing wells, for though wells in the positions of those in Figure 42 might get abundant water, they would not be likely to flow. If the stratum next below the aquifer is not impervious, some lower one probably is. No layer of rock is more impervious than one which is full of water, and the substructure of any bed which might serve as an aquifer is usually full of water, even when the rock, if free from water, would be porous.

If the outcrop of the aquifer is notably above the site of the well, and if it is kept full by frequent rains, the "head" will be strong, though the water at the well will not rise to the level of the outcrop of the aquifer. Experience has shown that an allowance of about one foot per mile of subterranean flow should be made. Thus if the site of the well is 100 miles from the outcrop of the water-bearing stratum, and 200 feet below it, the water will rise something like 100 feet above the surface of the well. This rule, is, however, not applicable everywhere. The failure of the water to rise to the level of its intake is due chiefly to the adhesive retardation of its movement through the rock. The more porous the rock, the less the retardation. The height of the flow is also influenced by the number of wells drawing on the same aquifer, on the degree of imperviousness of the confining bed above, etc. Flowing wells, often relatively shallow, are frequently obtained from unconsolidated drift.

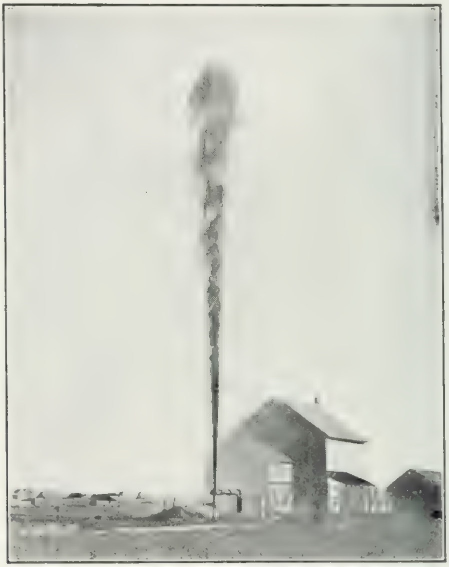

Fig. 43. — Artesian well at Woonsocket, S. D. (U. S. Geol. Surv.)

THE WORK OF GROUND-WATER

Ground-water is ever active. The results of its work are partly chemical and partly mechanical, the former being of greater importance.

THE CHEMICAL WORK

The results of the chemical and chemico-physical action of ground-water may be grouped in several more or less distinct categories.

1. Solution. The simplest effect is the solution of mineral matter. Pure water will dissolve certain minerals; but the carbon dioxide extracted from the atmosphere, and the products of organic decay extracted from the soil give ground-water a power to dissolve which is not possessed by pure water. The solvent work of ground-water is shown by the fact that all such water, whether it issues as springs or is drawn out through wells, contains mineral matter, from which rain-water is entirely free. The removal of soluble matter from rock makes it porous and causes decay.

Weathering. Where the solvent work of ground-water is slight and equally distributed, the result is to make the rock as a whole more porous. If, for example, some of the cement of sandstone is dissolved, the texture of the rock becomes more open; but if all the cement is removed, the rock is changed from a sandstone back to sand. If a complex crystalline rock contains among its many minerals one which is more soluble than the others, that one may be dissolved. This has the effect of breaking up the rock, since each mineral acts as a binder for the rest. It might happen that no one of the minerals is dissolved completely, but that some one of them is decomposed by water, and certain of its constituents removed. Such change would be likely to cause the mineral so affected to crumble, and with its crumbling, if it is an important constituent of the rock, the integrity of the whole is destroyed. The increase in volume

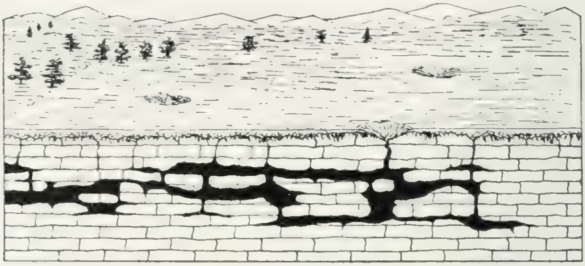

Fig. 44. — Diagram to illustrate the form and relations of caverns developed by solution. The black spaces represent caverns. Limestone sinks are represented at the surface.

attendent on hydration, oxidation, etc., sometimes leads to the disruption of the rock, as previously stated. These are phases of weathering.

Caverns. Where limestone is particularly soluble, groundwater may develop caverns and caves. One of the best-known regions of this kind is in the basin of the Ohio River in Kentucky and Indiana. Here caves are to be found literally by the hundred. While most of them are small, such caves as Mammoth and Wyandotte are miles in extent, and in places rather awe-inspiring in height.

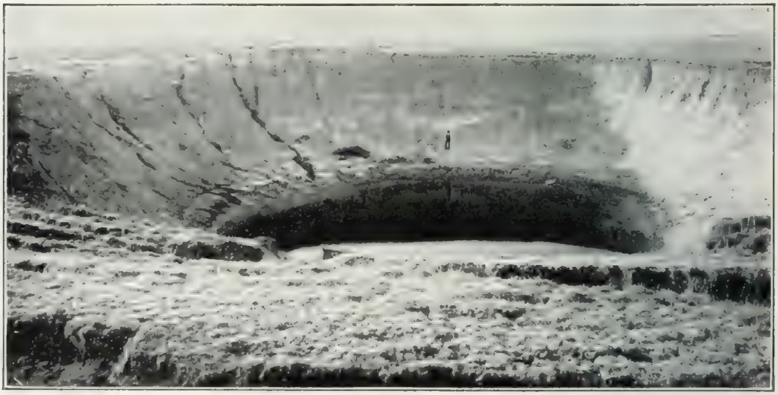

Fig. 45. — A sinkhole of recent development near Meade, Kan. (Johnson, U. S. Geol. Surv.)

Limestone Sinks. In regions of exceptionally soluble limestone the upper parts of cracks which have subterranean drainage may become funnel-shaped and greatly enlarged. Such depressions are *limestone sinks.* They may also be formed by the caving in of the roofs of caverns. Certain regions, as in parts of Kentucky, have these limestone sinks in such numbers as not only to dominate the topography but to control the drainage.

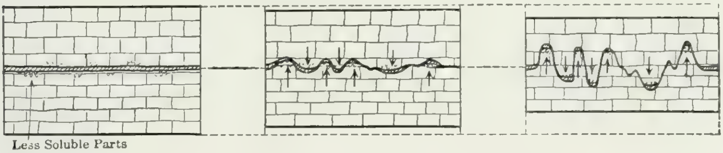

Fig. 46. — Diagram of stylolites.

Stylolites. Beds of rock are rarely of uniform solubility throughout. Consequently, ground-water, circulating along a bedding plane, in a soluble rock such as limestone, dolomite or marble (rarely quartzite) dissolves the more soluble parts of one bed, allowing the less soluble parts of the other bed in many cases to project as pointed teeth or columns.

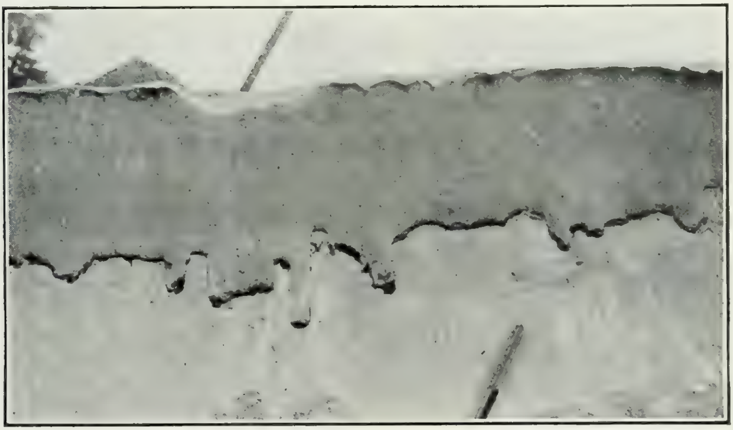

Fig. 47. — Stylolite seam, Salem limestone. Dark Hollow, St. Lawrence Co., Ind. (Stockdale.)

Solution is also augmented by increased pressure. Since the weight of the overlying beds rests more on the points of the projections than on their sides, solution is greatest opposite the projection, and the teeth will be caused to project farther and farther into the opposite bed. These interpenetrating columns or teeth alternately projecting upward and downward into the opposite bed are known as *stylolites.* The columns are generally capped by clay, which is left behind as the more soluble limestone is removed. There is thus a relation between the thickness of the clay cap, the length of the column, and the amount of insoluble impurity in the rock. The formation of stylolites involves the removal of much rock by ground-water. It is not unusual to find a score or more stylolites in a few score feet of strata, showing the removal of a considerable percentage of the original rock.

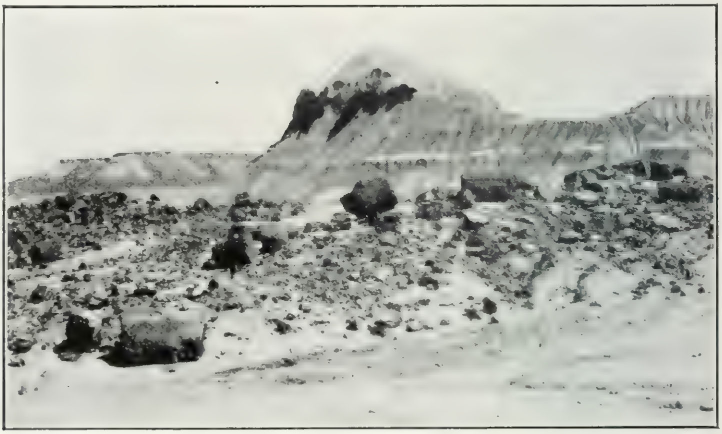

Fig. 48. — Petrified logs, near Adamana, Ariz. The logs have become exposed as the soft shale in which they occur has been washed away. (R. T. Chamberlin.)

2. Replacement. One mineral substance in a rock may be replaced by another from solution. Thus the lime carbonate of a shell embedded in rock may be removed, molecule by molecule, and some other substance, such as silica, left in its place. When the process is complete, the original substance of the shell has been completely removed, though its form and structure are preserved in the new material which has taken the place of the old. This process is known as *replacement.* Buried logs are, in places, converted into stone by the substitution of mineral matter for the vegetable tissue. This is *petrification* (Fig. 48). The minutest details of the structure are often preserved. Somewhat similarly, the molecules of one crystal may be replaced by a different mineral substance, e.g., the molecules of calcite by zinc carbonate. The zinc carbonate thus comes to have the natural shape of the calcite crystal instead of its own crystal shape. It thus has a false form and is called a *pseudomorph.* This sort of change may affect the crystals in almost any kind of rock.

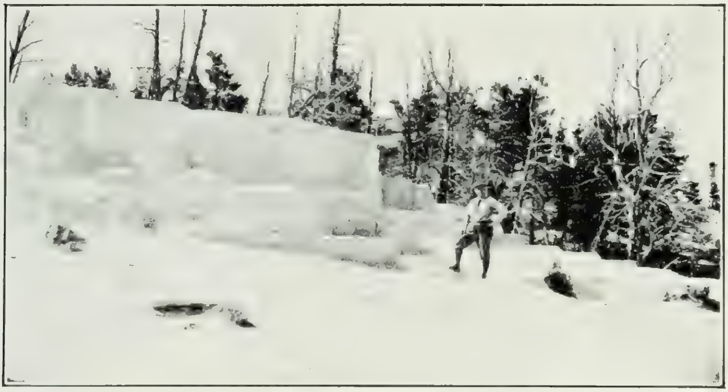

Fig. 49. — Hot spring terrace deposits. Angel Terrace. Yellowstone Park. (MacClintock.)

3. Deposition. Mineral matter is precipitated from solution in various ways. The most important of these are the following: (1) *If the water evaporates*, the dissolved mineral matter is left where the evaporation takes place. (2) *Reduction of temperature* often causes deposition. For this reason, ascending thermal waters as they cool deposit mineral matter in cracks below the surface or at the surface where they issue as hot springs. (3) Since for some substances such as limestone pressure increases the solvent power of water, *relief of pressure* may result in deposition. (4) In the mingling of certain unlike solutions *chemical reactions* take place, producing insoluble compounds which are precipitated. (5) If the water contains much gas, it has, in many cases, additional solvent power. *When the gas escapes,* as it does when pressure is relieved or the water is warmed, deposition results. (6) *Minute plants* (algae) thrive in many waters, and some of them cause the precipitation of mineral matter. This kind of deposit is seen about many springs, strikingly the hot springs in the Yellowstone Park.

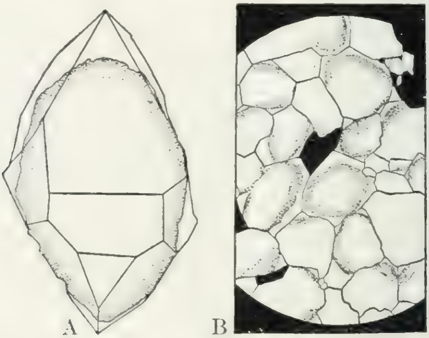

Fig. 50. — *A*, quartz crystal enlarged by secondary growth. The shaded outline represents the outline of the sand grain; the solid lines, the outline after secondary growth. Magnified 33 diameters. *B*, sandstone and quartzite texture. The shaded outlines represent the surfaces of the sand grains before growth, the intervening white portions, the added quartz, and the black portions, unfilled spaces. Open spaces characterize sandstone. When the spaces are filled with quartz, the rock becomes quartzite. Magnified 17 diameters. (Van Hise.)

Induration. The first step in the alteration of sediments is commonly their *induration*, largely by cementation. The cementing material which serves to bind together the grains of sedimentary rock is mineral matter deposited from solution in water. In many cases a striking plan is followed, as illustrated by the cementation of sand by silica. It is to be noted that grains of sand are more or less rounded particles of quartz, derived from rock which contains crystals of quartz. Interestingly enough, when the cementing quartz is deposited on these grains during induration, its molecules are oriented symmetrically with those of the sand grains, tending to build the grains out into the crystal form which quartz takes if free to grow in its own way. Microscopic examination of many well-cemented sandstones shows the original grains, distinct from the cementing material which appears as additions or secondary growths. The enlargement may go so far that the pores between grains are filled. Sandstone thus cemented becomes *quartzite* (Fig. 50). Between loose sand, at the one extreme, and quartzite at the other, there are all gradations. Grains of certain other minerals are subject to similar *secondary enlargement.* Important changes in rock are therefore brought about by solution and re-deposition of mineral matter through the influence of water in the rocks. This process of rock alteration might be called *aqueous metamorphism,* because of the important part played by water. Since water is nearly always present in rocks down to considerable depth, the changes which water produces are widespread.

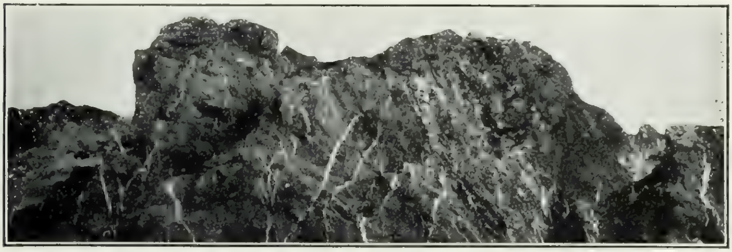

Fig. 51. — Veins of calcite in volcanic tuff. Shore west of Kincraig Point, Elie, Fife. (H. M. Geol. Surv.)

Cavity Filling. Cavities in rock larger than pores may also receive deposits, if the water entering them carries mineral matter in solution. The making of *veins* (Fig. 51), where a fissure is filled with mineral matter, is a case in point. Many veins of metallic ores arise in this way. The *agates* developed in some cavities afford another illustration. Here the successive layers are commonly silica (chalcedony) and differ from each other in color and texture. Before the cavity is entirely filled, the deposits may change from chalcedony to crystals of quartz which grow with their pyramidal points toward the center of the cavity. *Geodes* (Fig. 52) are examples of a similar process in which the cavity is but partially filled with crystals. The crystals of geodes are most commonly quartz or calcite, but they may be any other mineral which the waters are capable of carrying and depositing. Large cavities lined in this way are known to miners as *vugs*, and these grade into caverns lined with crystals. Among the most common forms of cave deposits are *stalactites* and *stalagmites* (Fig. 53). A stalactite may start from a drop of water leaking through the roof of a cave. Evaporation, or the escape of some of the carbon dioxide in solution in the water, results in the deposition of lime carbonate, about the margin of the drop, in the form of a ring. Successive drops make deposits on the lower edge of the ring, which grows downward into a hollow tube. Deposition in the tube may ultimately close it, while deposition from water

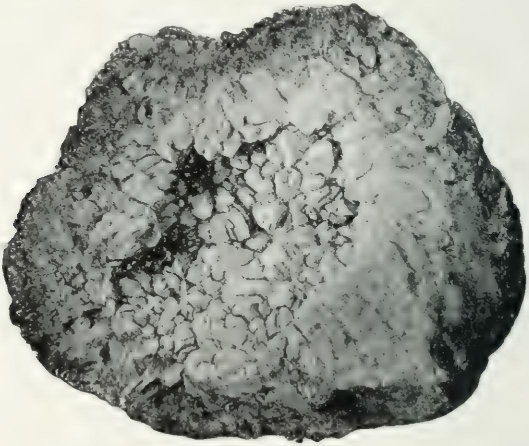

Fig. 52. — Geode. (Bassler, U. S. Geol. Surv.)

trickling down its outside may greatly enlarge it. By deposition from water dripping onto the floor, stalagmites may be built upward, even joining the stalactites to form pillars.

Concretions. Deposition about a nucleus in many cases forms a *nodule* or *concretion.* Ground-water may form concretions of calcium carbonate, common shale, of iron oxide, common in sandstone, or of silica (chert, flint), common in limestone.

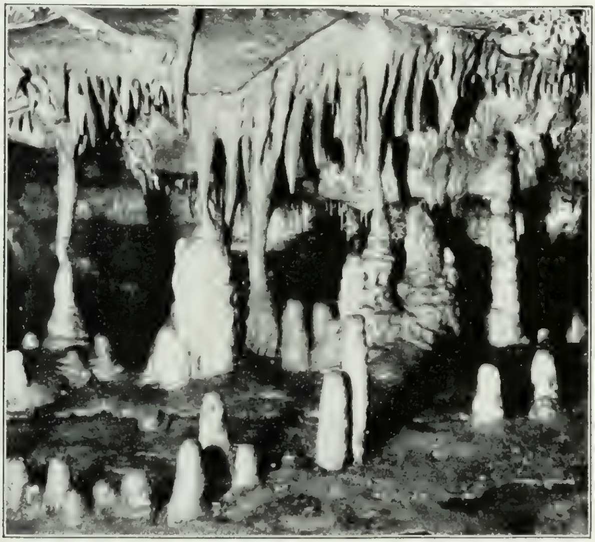

Fig. 53. — Deposits of calcite (travertine, stalactites, and stalagmites) in Wyandotte Cave, Ind. (Hains.)

Concretions, which range in size from microscopic dimensions to more than ten feet in diameter, show in some cases remarkable symmetry of form. Some concretions show internal cracks filled with vein material differing in color and composition from the main mass. Such concretions are called *septaria* (Fig. 57). In not a few cases the filling of the cracks appears to have wedged the segments apart, until the outer surface of the septarium is made up more largely of vein matter than of the original concretion.

Concretions are also formed by deposition in lakes or in the sea. In this case the precipitation takes the form of minute spherules which in aggregate so resemble the eggs of fish that they are called *oölites*. Oölitic limestone is common, while oölitic iron ore (flax-seed ore) is not rare. Calcareous oölites may subsequently be replaced by silica to form siliceous oölites.

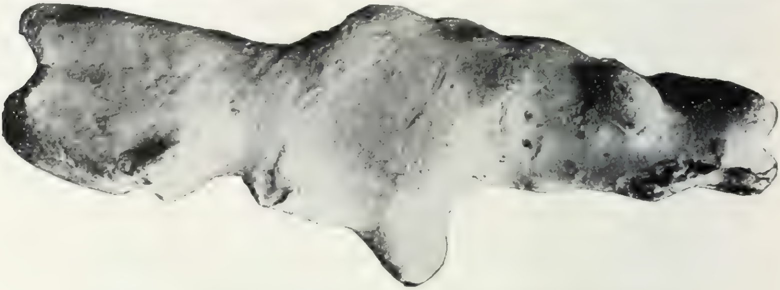

Fig. 54. — Nodule of chert, about half natural size. (Church.)

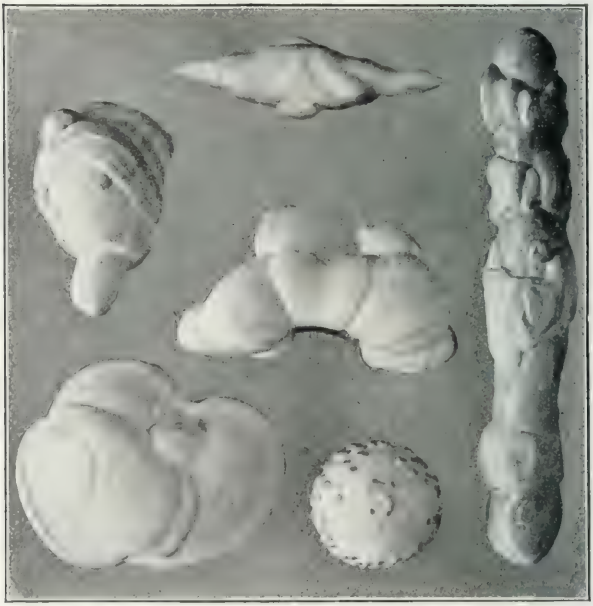

Fig. 55. — Irregular calcareous concretions. Ryegate, Vt. (Church.)

4. Combination. A further series of changes is effected by ground-water when it, or the mineral matter it contains,

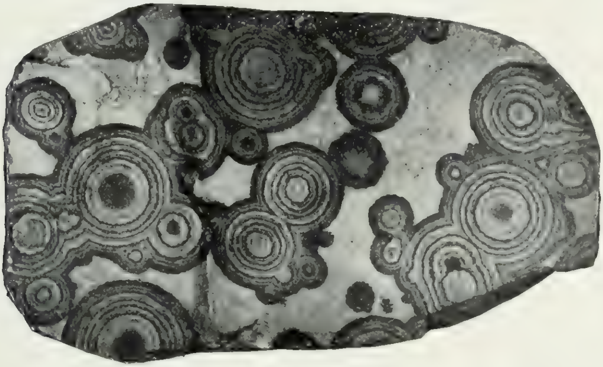

Fig. 56. — Section of spherical concretions, showing rings of ferruginous matter in sandstone; about $\frac{3}{8}$ natural size. (Geikie.)

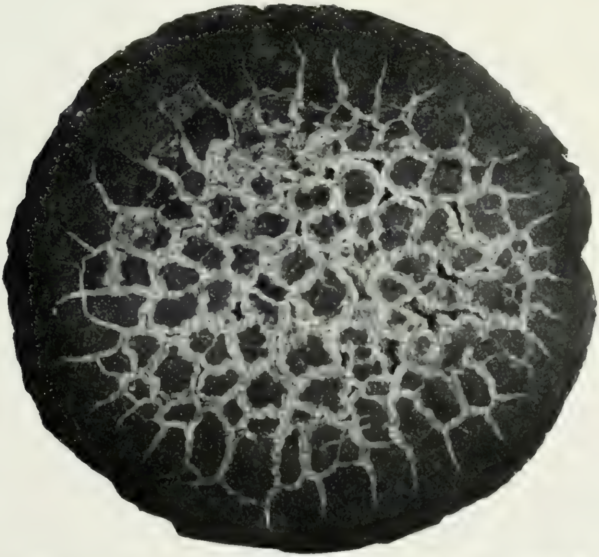

Fig. 57. — Section of a septarian nodule (clay ironstone). About $\frac{3}{8}$ natural size. (Geikie.)

enters into combination with the mineral matter through which it passes. One of the commonest processes of this sort, hydration, has already been referred to. But in the development of many of the commoner hydrous minerals, changes other than hydration are involved. These changes result in new *mineral combinations*, the new minerals being developed out of the old ones, usually with some additions or subtractions of material. In the long course of time changes of this sort may be so great as to alter rocks radically both chemically and physically.

GROUND-WATER AND ORE DEPOSITS

Many ore deposits are but a special result of the chemical work of ground-water. Metals are disseminated widely through the rock substance of the earth, but in their disseminated condition they cannot be considered as ore deposits. It is only by concentration of the metals into deposits from which they can be profitably extracted that ore bodies are formed. Many lead and zinc veins result from the solution of widely scattered lead and zinc compounds and their precipitation in veins of workable concentration. Some copper deposits owe their economic value to solution in the zone of aëration (oxidation) and precipitation of the copper minerals in the zone of saturation (*downward enrichment*). Ground-water also makes ore bodies by dissolving out the more soluble impurities, leaving the less soluble, desirable material in greater concentration than originally (residual concentration). In the Lake Superior region, silica has been dissolved out of the silicious iron formation, to leave the great deposits of nearly pure iron ore.

QUANTITATIVE IMPORTANCE OF SOLUTION

Calculations have been made which indicate, in a measure, the quantitative importance of the solution effected by ground-water, for the mineral matter in solution in streams is largely contributed to them by ground-water. In the case of several streams, among them the Thames and the Elbe, careful estimates of the amount of dissolved mineral matter have been made. The Thames drains an area only about one-tenth as large as the state of New York, but the amount of mineral matter that it daily carries in solution to the sea is estimated at 1500 tons. From the uppermost 20,000 square miles of its drainage basin, the Elbe is estimated to carry yearly about 1,370,000 tons of mineral matter in solution.

These figures make it clear that ground-water is an effective agent in the lowering of land surfaces. It is estimated that something like one-third as much matter is carried to the sea in solution as in the form of sediment, and that by this process of solution alone land areas of the globe are being lowered at the rate of about one foot in 30,000 years.[1]

The quantitative importance of the solution effected by ground-water is shown in another way. It is probable that most of the salt now in the sea has been taken to it in solution by waters flowing from the land. The amount of this salt is stupendous. Furthermore, most of the limestone of the earth has been extracted from sea-water, whither the largest part of it was carried by streams. The aggregate amount of limestone is far greater than the amount of salt in the sea. Some other sorts of rock, such as gypsum, of less importance quantitatively, have had a similar history. The total amount of rock which must have been decomposed by ground-water to yield such quantities of calcium carbonate, sodium chloride, etc., is far greater than *all the lithosphere now above sea-level.*

MECHANICAL WORK

The mechanical work of ground-water is so unimportant that it might almost be neglected. Very rarely is water below the surface concentrated into channels; and more rarely still is there enough current in the underground stream to pick up material or carry it off. Certain indirect effects, however, might be considered as partly due to the work of ground-water.

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[1] F. W. Clarke, *The Data of Geochemistry*, U. S. Geol. Survey Bull. 770 (1924), p. 121.

Creep, Slump, and Landslides. Alternate wetting and drying of the soil on steep slopes causes it to *creep* imperceptibly downwards. Ground-water may also so lubricate, as well as so increase the weight of surface material, as to cause it to *slump* or slide rapidly down steep slopes. Slumping is conspicuous along the sides of valleys, while landslides are very common on steep mountain slopes (Fig. 58). Landslides may be on so large a scale as to be destructive to life and property.

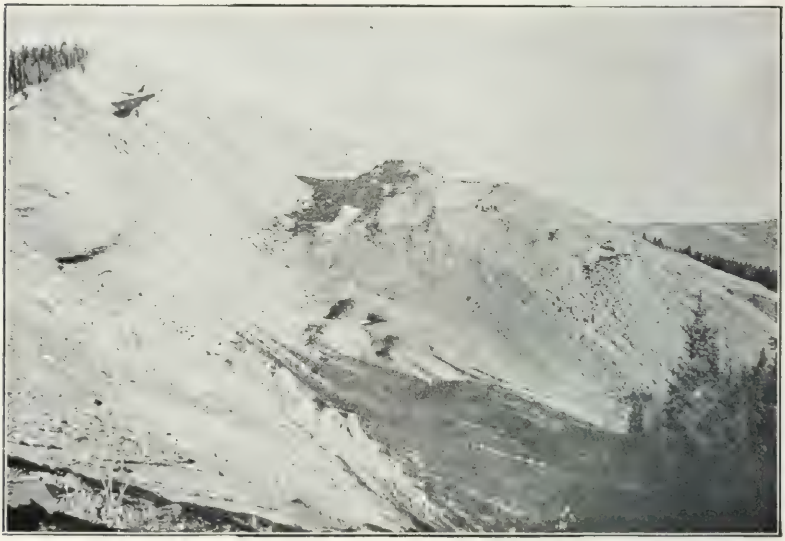


Fig. 58. — South face of Landslip Mountain, Colo. The protruding mass on the right has slumped down. (U. S. Geol. Surv.)

SUMMARY

All in all, ground-water is to be looked upon as a most important geological agent. When it is remembered that a large part of the water which falls on the surface of the earth, either in the form of rain or snow, sinks beneath the surface; that some of it sinks to great depths; that much of it has a long underground course before it reappears at the surface; that it is everywhere and always active, either in subtracting from the rock through which it passes, or in adding to it, in effecting the substitution of one mineral substance for another, or in bringing about new chemical combinations; and when it is remembered that these processes have been going on for millions of years, it will be seen that the total result accomplished must be great. The rock formations of the earth to which ground-water penetrates are to be looked upon as a sort of chemical laboratory through which waters are circulating, charged with many sorts of mineral substances. Some of the substances in solution are deposited beneath the surface, and some are brought to the surface where the waters issue. Much of the material brought to the surface in solution is carried to the sea and utilized by marine organisms in the making of shells. Without the mineral matter brought to the sea by springs and rivers, many shell-bearing animals of great geologic importance would perish. The importance of ground-water has obviously a biological, as well as a geological aspect.

CHAPTER IV

RUNNING WATER

Rivers are estimated to carry about 6500 cubic miles of water to the ocean annually. Since the average height of land is nearly half a mile, the waters which flow from the United States alone are estimated to develop 419,000,000 horse-power. The energy is therefore great, making them the great carriers of sediment from land to sea. The sediment which they carry is composed largely of decayed rock, but undecayed rock is also worn away, especially where the streams are swift. Though the flow of some streams is so gentle that they do not appear to work great changes in their valleys, others wear away their banks so rapidly that the changes which they produce may be seen from year to year or, when the stream is in flood, from day to day. The force of streams at such times is often disastrous. Occasionally they sweep away dams, bridges, and even buildings on their banks. The strong rods and beams of bridges, and

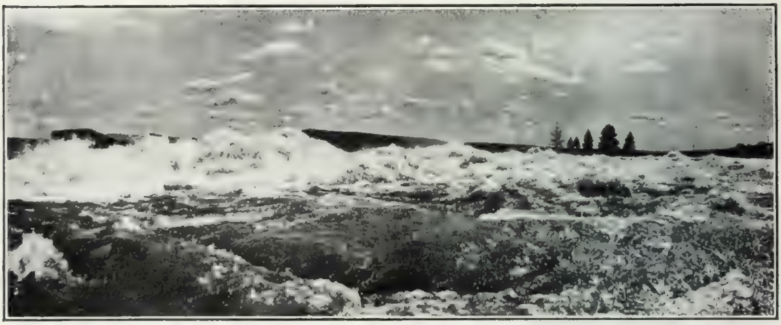


Fig. 59. — Spokane River, 4 miles above Spokane, during flood. (Tolman.)

the steel rails of railways are bent, almost as if they were twigs, by the force of the occasional torrent which follows an exceptional rain, as a cloud-burst (Fig. 60).

That the source of river-water is the rain and snow which fall from the atmosphere may be inferred from various familiar phenomena. Thus (1) streams are more numerous in regions where the rainfall is abundant than in those where it is scarce (Figs. 61 and 62); (2) multitudes of small streams spring into being with each heavy fall of rain and with each period of rapidly melting snow; (3) streams are notably swollen after rains, and most after heavy ones; and (4) many small streams, which flow during wet weather, dry up in times of drought, while practically all shrink. It is true that glaciers, lakes, and springs do feed rivers, but they themselves derive their supply of water from precipitation.

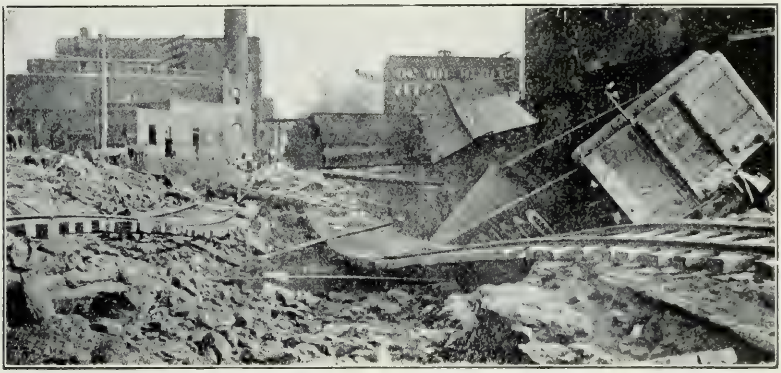


Fig. 60. — Scene in the freight-yards of Kansas City after the flood of 1903. (U. S. Weather Bureau.)

If the slope of a surface were perfectly even, the immediate run-off at any given time would flow in a sheet. There are slopes so smooth that water runs off them in this way; but on most slopes, even those which appear to be regular, there are small unevennesses, so that, although the run-off may start as a sheet, it is soon concentrated into rills and streamlets which follow the depressions. The smallest streamlets unite to form larger ones, and the little rills, after many unions with one another, reach the valleys which have *permanent streams.* These may be small (creeks or brooks) or large (rivers). Streams which flow but part of the time, as after a rainstorm, during wet weather, or during but part of the year, are *temporary* or *intermittent streams.*

Every permanent stream and many temporary ones flow in depressions called *valleys.* Valleys are therefore about as numerous as streams. The smaller depressions in which water runs only after showers are called *gullies* if they are very small (Fig. 63) or *ravines* if somewhat larger. Gullies and ravines are but small valleys, and just as the tiny streamlets unite to form creeks and these to form rivers, so the little gullies in which the smallest temporary streams flow, as a rule unite to form wider and deeper ones (Fig. 64). These, in turn, join one another and become ravines, which are but larger depressions of the same sort, and ravines lead to valleys just as gullies lead to ravines. Valleys, like streams, usually end at the ocean or a lake; but, in some cases, especially in arid regions, they end on dry land.

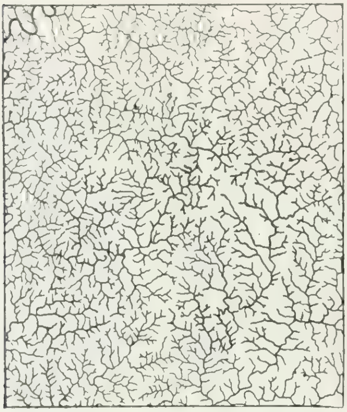


Fig. 61. — Map showing the many streams of a humid region. Central Kentucky. The area is about 225 square miles.

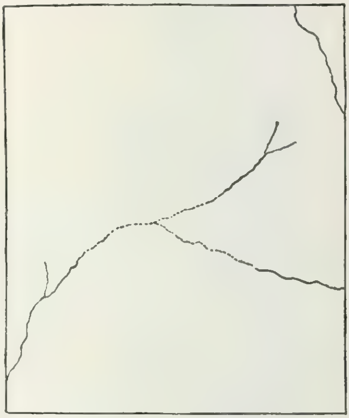


Fig. 62. — Map showing the few streams of an arid region. Northern Arizona. The area is as great as that shown in Fig. 61.

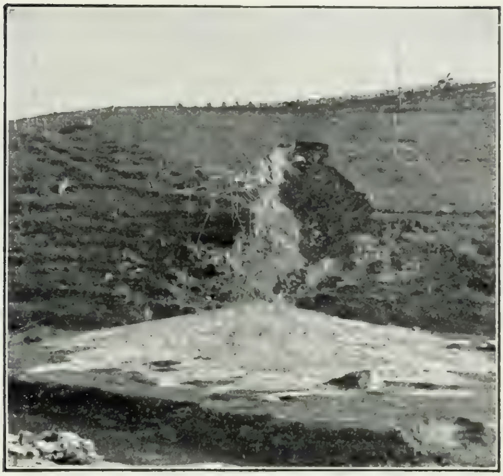


Fig. 63. — A gully developed by a single shower. (Blackwelder.)

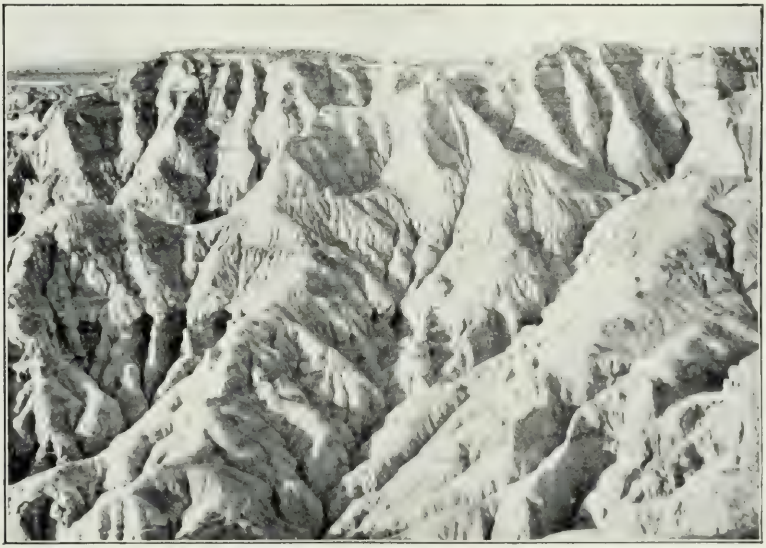


Fig. 64. — Slope with numerous gullies, the smaller ones joining the larger ones. Scott's Bluff, Neb. (U. S. Geol. Surv.)

There is, as a rule, a relation between the size of the valley and the stream which follows it, though this relation is not one which can be stated in mathematical terms. The large stream and the large valley go together so often, however, that the combination cannot be accidental.

EROSION BY RUNNING WATER

Analysis of Erosion. Erosion is the term applied to all the processes by which rock (mantle or solid) is loosened, picked up, and carried away. In its application to running water it signifies the *wearing away of the land.* Erosion consists of three more or less distinct sub-processes: (1) *weathering,* (2) *corrasion,* and (3) *transportation.*

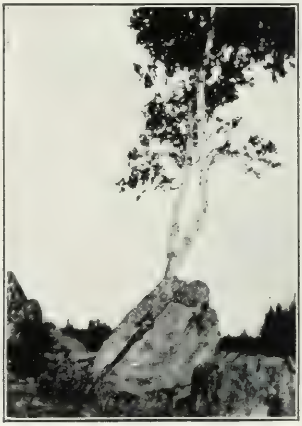


Fig. 65. — Tree growing in crack in a rock, and by its growth splitting the rock.

WEATHERING

As defined earlier, weathering includes all those unobtrusive processes by which rock is disintegrated and disrupted. The lettering of inscriptions on exposed marble becomes fainter as time goes by and finally disappears, because the rock in which the letters were cut has been weathered away. In this case, the weathering is effected partly by solution in rain-water, but largely by work of the atmosphere.

The rain which falls upon the surface of exposed rock, and that which sinks through the soil to the solid rock below, dissolves slowly some of the constituents of the rock. This tends to make the rock crumble, much as mortar does when the lime carbonate which cements the sand is dissolved. The chemical changes effected by ground-water and its dissolved gases also help, as we have seen, to disintegrate the rock.

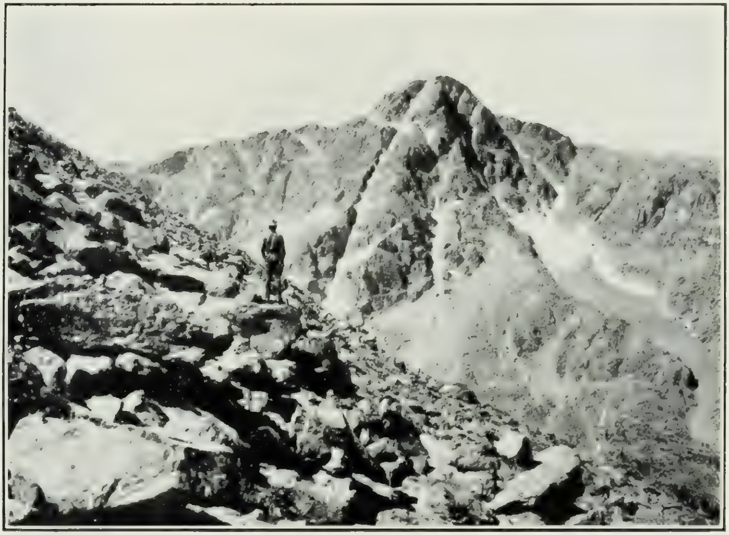


Fig. 66. — Talus slopes in the mountains. Mount of the Holy Cross, Colo. (Courtesy U. S. Forestry Serv.)

There are other processes of weathering not due directly either to the atmosphere or to water. The roots of trees frequently grow into cracks of rocks (Fig. 65) and, increasing in size, act like wedges; water freezing in cracks works the same way.

From the faces of steep cliffs, masses of rock are frequently loosened, by the wedge-work of roots or ice, or by expansion and contraction due to changes in temperature. The quantities of debris at the bases of many cliffs, forming slopes of talus (Fig. 66), testify to the importance of the weathering of rock, and to the action of gravity in getting loosened material down. Another phase of gravity work is shown in Figure 67. Here, under the influence of gravity and expansion and contraction, due to the freezing and thawing and wetting and drying, the surface material is creeping down slope. In the process the rock is being broken. The descent of detritus under the influence of gravity is hardly weathering, but it is often weathered material which is brought down.

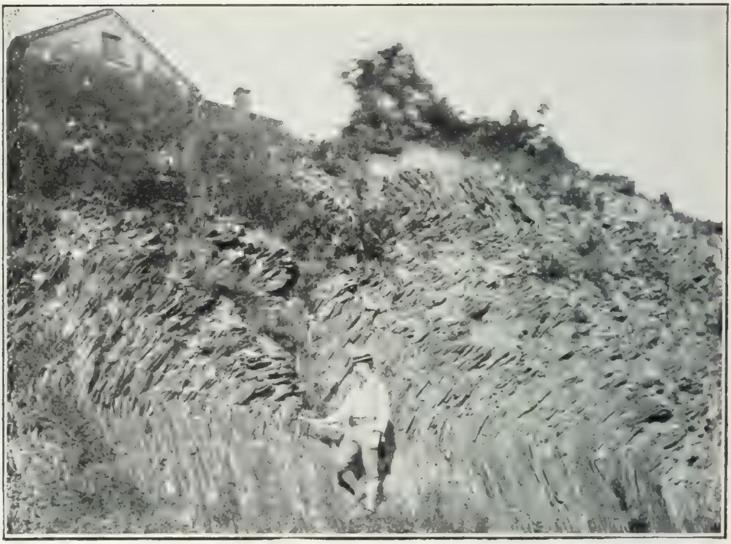


Fig. 67. — Shows the downward creep of soil and slaty rock under the influence of gravity. (U. S. Geol. Surv.)

The importance of weathering in the general processes of erosion is shown in many ways. In regions where the mantle rock is the product of the decay of the solid rock beneath, and such regions constitute a large portion of the earth's surface, this mantle rock represents the excess of weathering over transportation. Since most of the earth's surface is covered with mantle rock, it is clear that, on the whole, weathering keeps ahead of transportation. Again, it is clear that the loosening of the rock by weathering greatly increases the erosion which a given amount of moving water can accomplish. Not only this, but weathering plays a much more important rôle in the development of valleys than is commonly realized. This is illustrated by the valleys of swift streams. The valley whose top is not ten times as wide as its stream, is rare. The stream which has such a canyon has been cutting chiefly at its bottom. Weathering, slope wash, etc., have widened the valley especially in its upper part, (Fig. 68), and the stream has carried away the material brought down to it. The foregoing illustration would not apply to sluggish streams for they widen their valleys, by meandering, independently of weathering.

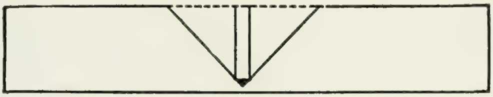


Fig. 68. — Diagram of a valley, the top of which is ten times the width of the stream.

The central idea of the processes embraced under the term weathering is the loosening and the disrupting of rock, by which it is prepared for transportation. Weathering is the first step in most erosion, but it is not the only one, and erosion may take place without it.

CORRASION

The mechanical wear effected by running water is *corrasion*. It is easy to understand how running water moves materials that are incoherent. As water flows over the surface of a cultivated field it gathers earthy matter, and the process is continued all the way to the channel of the stream. Thus sediment is gathered at the very sources of the streams.

It is also contributed by all slope wash from whatever part of the valley, and the stream gathers more load from its bed wherever it flows with sufficient velocity over loose material. Streams also undercut their banks, and receive new load from the fall of the overhanging material.

By far the larger part of the sediment of most streams is made up of material loosened in advance by the processes of weathering. The stream, and the waters which get together to make the stream, find this material ready-made; but rivers frequently wear rock which is not weathered, for the principal valleys of the earth are in solid rock, and many of them in rock of great hardness. How does a stream wear solid rock?

When a stream flows over a bed of solid rock, the wear which it accomplishes depends chiefly on the character of the rock, the velocity of the stream, and the load it carries. If the rock is much divided by bedding planes and joint planes, the water of a clear stream of even moderate strength may dislodge bits of the rock. This condition of things is often seen where streams run on beds of shale or slate. If the rock is hard and without bedding planes and joints, or if its layers are thick and its joints few, clear water will be much less effective. If massive hard rock presents a smooth surface to a clear stream, the mechanical effect of even a swift current is slight.

This general principle is illustrated by the Niagara River. Just above the falls the current is swift and the water clear. The surface of the limestone bed is seen to be distinctly green from the presence of small plants (fresh water algae), which the whole force of the mighty torrent is not able to sweep away. Were the stream supplied with a tithe of the sand which it is capable of carrying, it would not take long to remove the last trace of vegetation. This illustration furnishes a clue to the method by which the erosion of solid rock in a stream's bed is accomplished.

The gravel rolled along the channel wears even solid rock, and as the moving stones wear the stream's bed, they are themselves worn by impact with the bed and with one another, and are reduced to rounded water-worn forms. The particles broken off may make grains of sand, or if very fine, particles of silt or mud. In the course of time the pebbles or cobbles rolled along may be literally worn out.

The sediment carried in suspension, as well as that rolled along the bottom, wears the bed of the stream. The coarser the sediment and the stronger the current, the greater the wear. Other processes are largely concerned in making the valleys wide, but the depth of the valleys cut in solid rock is chiefly the result of the impact and friction of the sediment in transportation. The gravel, sand, and mud carried by the stream are therefore tools with which it works. Without them it is relatively impotent, as far as the abrasion of solid rock is concerned; with them, the stream may wear any rock over which it passes.

TRANSPORTATION

A third element of erosion is transportation. The transportation of sediment is to be distinguished from the transportation of materials in solution. In so far as mineral matter is dissolved it becomes a part of the fluid of the stream. If the quantity dissolved were large, it might influence the mobility of the water; but the amount is usually too slight to influence flow sensibly.

The sediment transported by the stream, whether rolled along its bottom, or carried in suspension above the bottom, is its *load*. The coarser materials (gravel and sand) are carried chiefly in the former position and the finer (silt and mud) largely in the latter.

Transporting Power and Velocity. The transporting power of running water depends on its velocity. Swift streams have enormously greater power of transportation than sluggish ones, but transportation does not always keep pace with transporting power. The Niagara at its rapids is a stream of great transporting power, but it carries little sediment, because there is little to be had.

The velocity of the stream depends on three elements: its gradient (slope), its volume, and its load. The higher the gradient, the greater the volume, and the less the load, the greater the velocity. The relation between gradient and velocity is evident; that between velocity and volume is illustrated by every stream in time of flood, when its flow is greatly accelerated. The relation between velocity and load is less obvious, but none the less definite. Every particle of sediment carried by the stream makes a draft on its energy, and energy expended in the carrying of the load reduces the velocity. A muddy stream is never so swift as a clear stream of the same size would be in the same channel.

How Sediment is Carried. Coarse materials, such as gravel and cobbles, are rolled along the bottom of the swift streams, their movement being due to the impact of the water. The same is true to a large extent of sand grains. So far as concerns the material rolled along the bottom, it is to be noted that a stream's transporting power is dependent on the velocity of the water at its bottom. This is much less than the surface velocity, or even the average velocity.

The particles of fine sediment, such as silt and mud, are commonly carried by streams quite above their bottoms, as shown by the muddiness of many streams. Most particles of mud are small bits of mineral matter, the specific gravity of which is between two and three times that of the water. Why do they not sink through the water and come to rest on the bottom?

A particle of sediment in running water is subject to two principal forces, that of the current which tends to move it nearly horizontally downstream, and that of gravity which tends to carry it to the bed of the stream. As a result of these two forces, the particle tends to move in the general direction which represents the resultant of the forces (Fig. 69). If a river were the simple straight-forward current which it is popularly supposed to be, a particle in suspension would reach its bottom in the time it would take to sink through an equal depth of still water, for the descent would be none the less certain or prompt because of the forward movement of the water. The current would simply be a factor in determining the position of the particle when it reached the bottom, not affecting the time of reaching it. Very fine particles, like those of clay, have great surface area relative to their weight. They consequently offer more friction in sinking than do coarse particles, and so sink more slowly. But even such particles, unless of extraordinary fineness, would presently reach the bottom if acted on only by horizontal current and gravity. Since much sediment, not of exceeding fineness, is kept in suspension, it is clear that some other factor is involved. This is found, in part at least, in the subordinate upward currents of the stream.

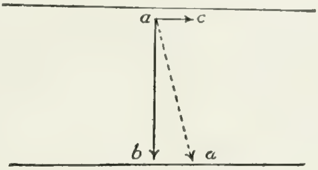


Fig. 69. — Diagram to illustrate the relative strength of the two forces acting on a particle in suspension. The arrows represented by full lines show the relative strength of the two forces when the stream's velocity is about 5 miles per hour. No account is taken in the diagram of the viscosity of the water, or of the acceleration of velocity of fall.

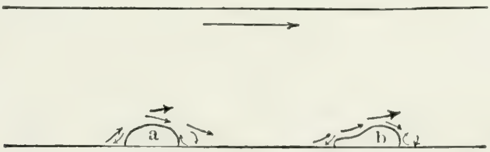


Fig. 70. — Diagram to illustrate the effect of irregularities, *a* and *b*, in a stream's bed, on the current striking them.

Where a boulder lies in the bed of a stream (Fig. 70) a part of the water which strikes it is deflected upward. If there are many boulders, the process is repeated in many places, and the number of upward currents is great. Any roughness will serve the same purpose, and every stream's bed is rough to a greater or less extent. Roughness at the sides of a channel starts currents which flow toward the center of the stream, and the varying velocities of different parts of the stream serve a similar purpose. A river is therefore to be looked upon as a multitude of currents, some rising from the bottom toward the top, some descending from top to bottom, some diverging from the center toward the sides, and some converging from the sides toward the center.

It is, of course, true that the sum of the upward currents is always less than the sum of the downward, so that the aggregate motion of the water is down slope. Sediment in suspension is kept up chiefly by the upward currents, which, locally and temporarily, overcome the effect of gravity. The particles in suspension are constantly tending to fall, and frequently are falling; but before they reach the bottom, many of them are seized and carried upward by the subordinate currents, only to sink again and be carried up again. Even if they reach the bottom, as they frequently do, they may be picked up again. It is probable that every particle of sediment of such size that it would sink readily in still water is dropped and picked up many times in the course of any long river journey, and its periods of rest probably often exceed its periods of movement.

VALLEY DEVELOPMENT

Beginning of a Valley. 1. If water flows in a sheet down a uniformly sloping surface, *sheet erosion* results. But if the slope of the surface is not uniform, the effect is very different. If there is, for example, a slight depression near the base of the slope, more of the descending water flows through it than over other parts of the surface, and the greater volume of water following the depression gives it greater velocity. Greater velocity causes greater erosion, and greater erosion deepens the depression. The immediate result is a gully or *wash.* As soon as the gully is started, it concentrates drainage still more, and it is thereby enlarged. The water which enters from the sides widens it; that which enters at the head lengthens it by causing its upper end to advance up the slope; and all which flows through it deepens it. The enlarged gully will gather more water to itself, and as before, increased volume means increased velocity, and increased velocity, increased erosion. As the gully grows, therefore, its increased size becomes the occasion of still further growth; continual growth transforms the gully into a ravine, though there is no distinct line of demarcation between the two. But growth does not stop with ravinehood. Water from every shower gathers in the ravine, and

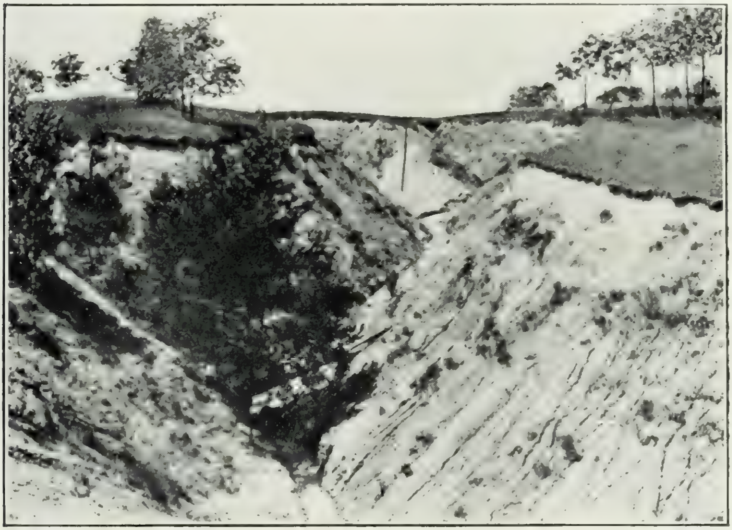


Fig. 71. — A young gully. (Senstius.)

flowing through it, increases its length, width, and depth, until it becomes a valley.

It was assumed in the preceding paragraphs that the single depression in the slope was meridional and low on the slope; but almost any sort of depression in almost any position would bring about a similar result, since it would lead to concentration of the run-off. Had the original surface been marked by a ridge instead of a depression, valley development would have been much the same, for a ridge, like a depression, would cause the concentration of the run-off along the sides or around the ends, and therefore lead to the development of valleys.

Under the conditions represented in Figure 72 the lengthening of the drainage depression is affected chiefly at its upper end, the head of the valley working its way farther and farther back. This method of lengthening is known as *headward erosion.* But the lengthening of the valley is not always wholly by headward erosion. The gully begins normally where concentration of run-off begins, and if this is not at sea-level, the gully may be lengthening at both ends at the same time. This would have been the case, for example, had the original depression of Figure 72 been half way up the slope of the island. In the case of land emerging from the sea, coastal streams may lengthen their valleys seaward with the retreating shore. Valleys developed under the control of surface slope are *consequent valleys*, and their streams are *consequent streams.*

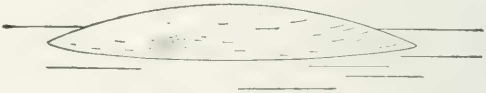


Fig. 72. — Diagram showing a slight meridional depression in the surface of an otherwise even-sloped island.

2. If the surface material of a slope is of unequal resistance, the water flowing over it will develop irregularities of slope, even if the slope was uniform at the outset. Thus

if the material of a certain part of the slope is less resistant than that elsewhere, the run-off will erode most in this part. The depression thus started will be a cause of its own growth, and, as before, the gully may develop into a valley.

If there be sufficient rainfall, therefore, either heterogeneity of *slope*, or of *material*, will occasion the development of valleys. If lack of uniformity appears at but few points, there will be but few valleys; if at many points, the number of valleys will be large. Since it is unlikely that any great area of land ever had perfectly homogeneous material and absolutely uniform slopes, every considerable land area, affected for any considerable length of time by abundant rain, has had valleys developed in it, and the heterogeneity of material and slope is usually such that valleys are developed at short intervals.

The Permanent Stream. It appears from the foregoing discussion that a valley may be developed by the run-off of successive showers. If supplied from this source only, surface streams would cease to flow soon after the rain stopped, and a valley might attain considerable size without possessing a permanent stream. How does a valley developed by the run-off of successive showers come to have a permanent stream?

When a valley has been deepened until its bottom is below the water-table, water seeps or flows into it from the sides. The valley is then no longer dependent upon the run-off of showers for a stream. When the bottom of a valley is below the water-table of a wet season without being below that of a dry season, it will have an intermittent stream. If the rainfall of the year were concentrated in a single wet season, the intermittent stream would flow not only during that season, but for so long a time afterward as the water-table remained well above the valley bottom. In regions subject to frequent and short periods of precipitation alternating with droughts, the periods of intermittent flow may be many and short. Many valleys are now in the stage of development where their streams are intermittent.

As a valley containing an intermittent stream becomes deeper, the periods when it is dry become shorter, and when it has sunk below the lowest water-table, it will have a permanent stream (3, Fig. 73). Since a valley normally develops headward, its lower and older portion is likely to acquire a permanent stream while its upper and younger part has only an intermittent one. So soon as a valley gets a permanent stream, the process of valley-enlargement goes on without the interruption to which it was subject when the supply of water was intermittent. In general a permanent stream at one point in a valley means a continuous stream from that point to the sea or lake which the valley joins; but to this rule there are many exceptions. They are likely to arise where a stream heads in a region of abundant precipitation, and flows thence through an arid tract where the water-table is low, and evaporation is great. In such cases evaporation and absorption may dissipate the water gathered above, and the stream disappear (Pl. II).

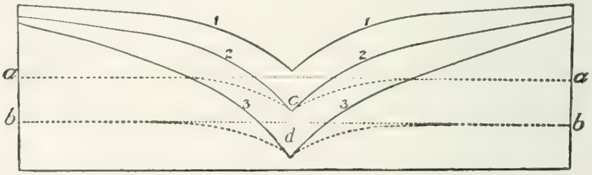
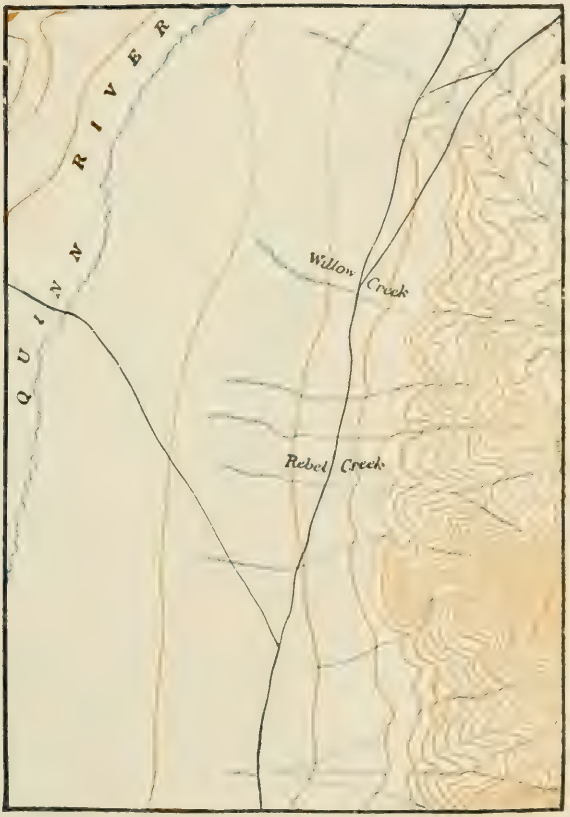


Fig. 73. — Diagram to illustrate the intermittency of streams due to fluctuations of the ground-water level. The water-level *aa* would be depressed next the valley 2-2, by the flow of the water into the valley. The profile of the ground-water surface would therefore be *aca* rather than *aa*.

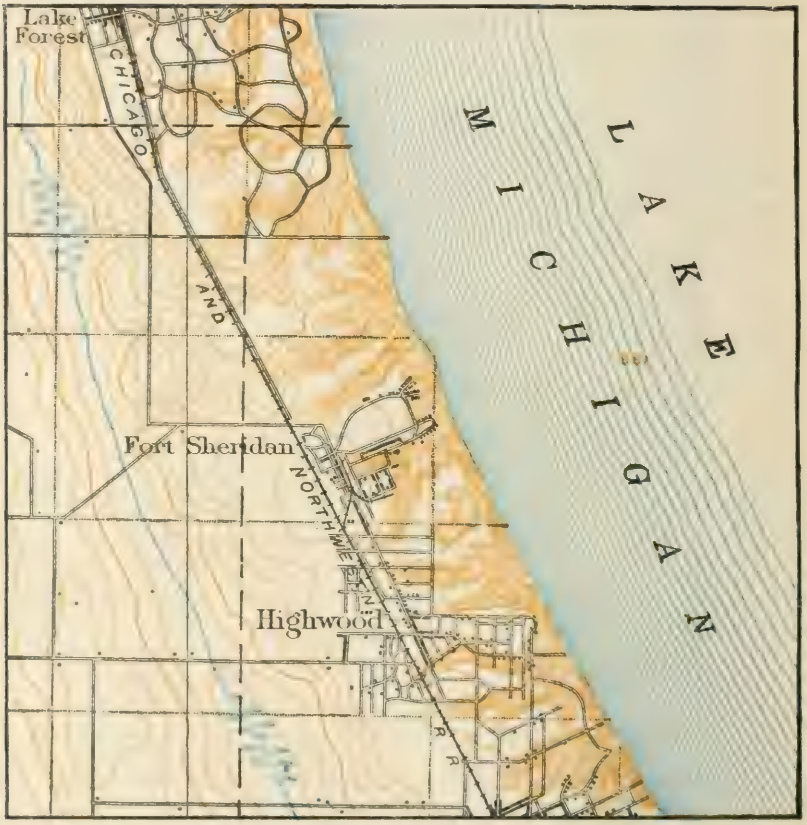
Other Modes of Valley Development. Valleys are not always developed from gullies in the manner outlined above. The out-flow of a lake would develop a valley, and the valley might be in process of excavation all the way from the lake basin to the sea at the same time. A valley developed in this manner is not simply a gully grown big by

PLATE II


QUINN RIVER
Willow Creek
Rebel Creek

Streams disappearing in the sand, gravel, etc., at the base of mountains in an arid region. Scale, about 4 miles per inch. (Paradise, Nev., Sheet, U. S Geol. Surv.)

PLATE III


Lake Forest
CHICAGO
AND
NORTHWE
N
R R
Fort Sheridan
Highwood
LAKE
MICHIGAN

Youthful valleys. Shore of Lake Michigan just north of Chicago. Scale, about 1 mile per inch. (Highwood Sheet, U. S. Geol. Surv.)

headward erosion, and the valley would not precede the permanent stream.

If a narrow coastal plain is limited landward by a steeper slope, valleys might develop as shown in Figures 74 and 75.

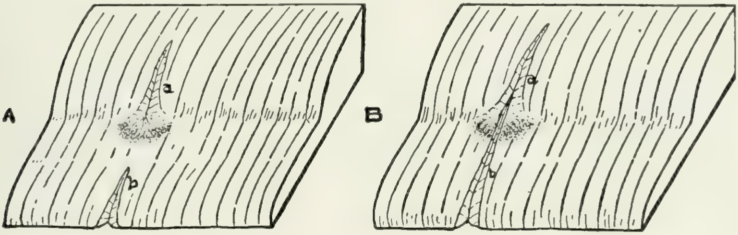


Fig. 74

Fig. 75

Figs. 74 and 75. — Diagrams to illustrate one mode of valley lengthening. In Fig. 74 there are two small valleys, *a* and *b*, and the former ends at the base of the steep slope. In Fig. 75 valley *b* is represented as having been lengthened so as to join *a*, and the two have become one.

Again, in mountain regions valleys are in places formed by the uplift of parallel mountain folds, which leave depressions between (Fig. 76). Drainage will appropriate such valleys so that they become in some sense river valleys. But they are not river valleys in the sense in which the term has been used in the preceding pages. They are rather *structural valleys.* River valleys may be developed in their bottoms (a, Fig. 76) and they may be in process of development through the whole length of the structural valleys at the same time.

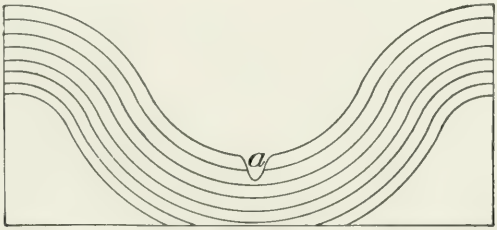


Fig. 76. — Structural valley with a river valley developing in its bottom.

These illustrations do not exhaust the list of conditions under which valleys develop, but they suffice to show that valleys originate and develop in different ways.

LIMITS OF VALLEY GROWTH

1. Depth. Streams which are swift because of high gradient erode their channels downward; but by this downward erosion the slope is made progressively less and less. It follows that a stream, cutting its valley lower and lower, becomes more sluggish and hence less able to erode downward. Figure 77 illustrates this principle. The stream is steep from *a* to *s*. It grows less steep as it cuts to *bs*, *ds*, and *es*. It is apparent that the slope *fs* is an *erosion base* below which the stream does not cut its valley. It is the lowest level to which the stream can cut its valley by mechanical wear, and is the *level of base-level.* But widening of the valley may continue, and eventually the stream and its tributaries may reduce all the surrounding country to this common level, or base. *Any considerable area of land reduced to essential flatness by the erosion of running water is a base-level.*

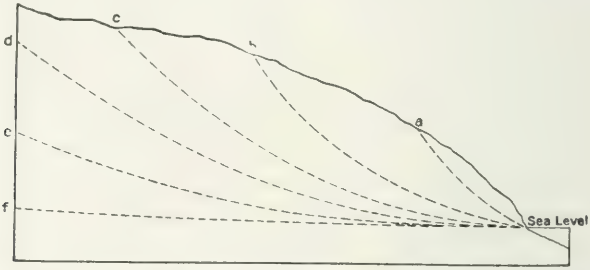


Fig. 77. – Diagram of depth limit of erosion by running water.

A clear distinction should be made between the *level of base-level* and a *base-level.* The former is an imaginary plane lying below the surface of all unreduced land, while the latter is an actual land surface. A comparison may illustrate the point. Before a road or street is built plans are drawn in the engineer's office. In these plans the *level* of the road is specified. This imaginary level of the future road corresponds to the *level of base-level.* Surveyors then lay out

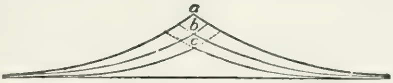


Fig. 78. — Diagram to illustrate the lowering of a divide without shifting it. The crest of the divide is at *a*, *b*, and *c* successively. If the erosion was unequal on the two sides, the divide would be shifted.

the road and it is built. The actual surface of the completed road corresponds to a *base-level.*

2. Length. Valleys are lengthened by headward erosion till a permanent *divide* is established. If the divide between the headwaters of two streams is eroded more rapidly on one side than on the other, it is not permanent but will be shifted, as it is lowered, away from the more powerful stream. It follows that a permanent divide is one which *suffers erosion at the same rate* on the two sides, and therefore is not shifted in position as it becomes lower and lower. (Fig. 78).

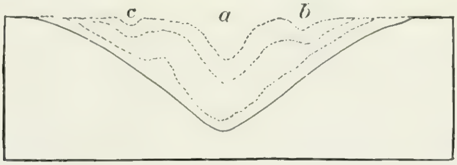


Fig. 79. — Diagram illustrating how one gully absorbs another as a result of lateral erosion.

3. Width. The widening of valleys is limited much as is their lengthening. Adjacent valleys grow wider until the tops of the intervening divides are reduced to narrow crests. Then, if erosion is equal on the two sides, the divides are permanent and are lowered without shifting, so far as slope wash is concerned.

DEVELOPMENT OF TRIBUTARIES

Most considerable valleys have numerous tributaries. As soon as a gully is started, the water flowing into it down its sides wears back the slopes. Slight inequalities of slope, or variations in the character of the material, are sufficient to make the erosion on the slopes unequal at different points, and unequal erosion on the slopes results in the development of tributary gullies. Tributary gullies develop into ravines and valleys, the same as their mains. Every new valley

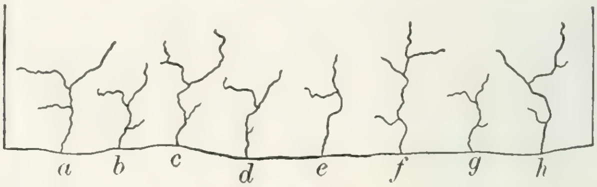


Fig. 80

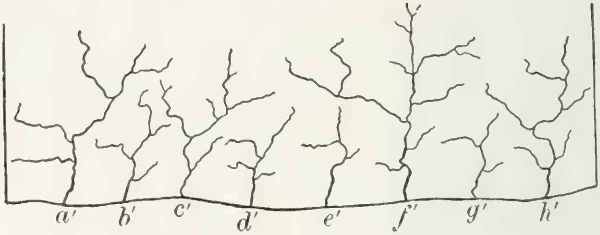


Fig. 81

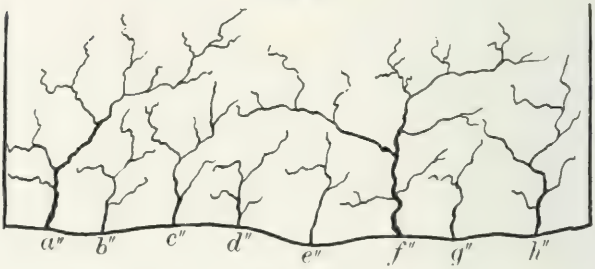


Fig. 82

Figs. 80, 81, and 82. — Diagrams illustrating successive stages in the struggle for existence among streams.

facilitates the run-off of the water which falls on the land, and so helps along erosion.

THE STRUGGLE FOR EXISTENCE AMONG VALLEYS AND STREAMS

It is not to be inferred that every gully becomes a valley, nor that every small valley becomes a large one. The number of miniature valleys which develop on a slope may be very large (Fig. 64); but the history of many of them is ephemeral. If two adjacent ones are of unequal depth, the widening of the deeper one narrows and finally eliminates the divide between them, and the two become one (Fig. 79).

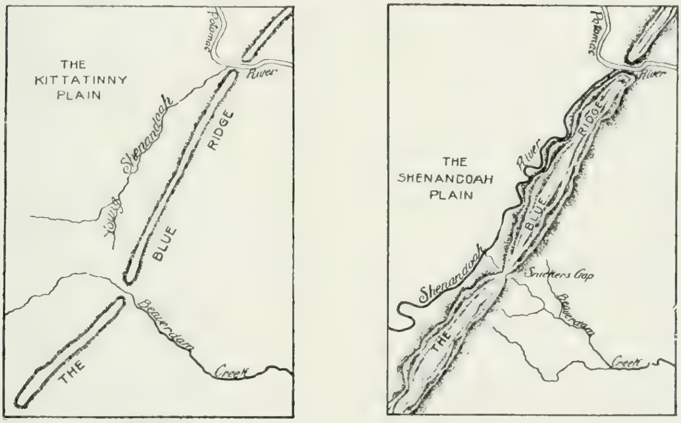

THE KITTATINNY PLAIN
Potomac
River
Shenandoah
Young
BLUE RIDGE
THE
Beaverdam
Creek
Potomac
River
THE SHENANDOAH PLAIN
River
THE BLUE RIDGE
Shenandoah
Snickers Gap
Beaverdam
Creek

Fig. 83

Fig. 84

Figs. 83 and 84. — The capture of the head of Beaverdam Creek by the Shenandoah River. Virginia-West Virginia. (After Willis.)

Any good map of the north shore of Lake Superior, or the west shore of Lake Michigan, shows a large number of small streams and gullies (Pl. III). No equal stretch of coast has a number of large valleys comparable to that of the small ones along such shores. It therefore seems evident that of these many small valleys a few only will attain considerable size.

Some of the young valleys work their heads into the land faster than others, because of inequalities of slope and material. If valleys develop in ways other than by head erosion, the chances are also against their equal growth. If two streams, such as *a* and *c* (Fig. 80), develop faster than the intermediate stream *b*, it is clear that their tributaries may work back into the territory which at the outset drained into *b* so as to cut off the supply of water from the latter stream (compare *a′*, *b′*, *c′*, Fig. 81). As a result, the growth of *b* will be checked and ultimately stopped. Similarly other valleys, such as *f* (Fig. 80), will get the better of their neighbors, and many of the competitors, as *d′*, *e′*, and *g′* (Fig. 81) will soon drop out of the race. Between the stronger streams competition still goes on. If *a′* and *f′* (Fig. 81) develop faster than *c′*, the prospective drainage territory of *c′* will be pre-empted by its rivals (compare Figs. 81, 82). It is apparent that as a result of the unequal

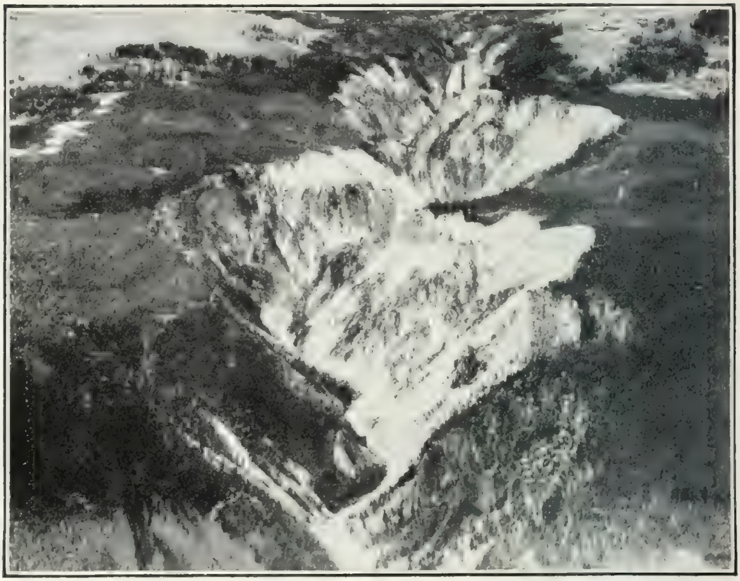


Fig. 85. — Young valley. Canyon of the Yellowstone.
(Courtesy U. S. Air Serv.)

rate at which the valleys are lengthened, the larger number of those which come into existence are arrested in their development.

Piracy. Streams do not always hold the courses which they establish for themselves in youth. Thus the valley of the strong Potomac River across the Blue Ridge was (Fig. 83) deepened faster than that of the smaller Beaverdam Creek. The divide between the head of the young Shenandoah River and the Beaverdam was consequently not permanent, for it was eroded more on the north by the river having a steeper gradient down to a lower gap through the Blue Ridge. It followed that the divide was shifted to the south by the headward erosion of the young Shenandoah till the upper waters of Beaverdam Creek were diverted (Fig. 84). The young Shenandoah was the *pirate*, and Beaverdam Creek was *beheaded*. The abandoned course of the creek across the ridge is a *wind-gap*.

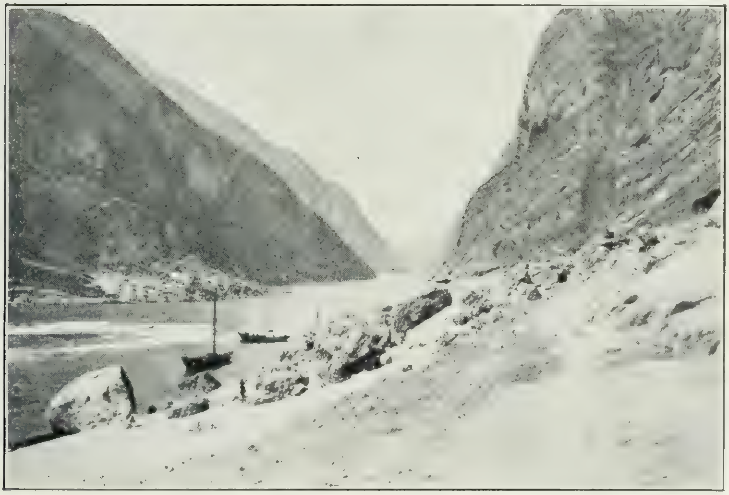


Fig. 86. — Large valley in youth. Sintan Rapids. Gorges of the Yangtze Kiang, China. (R. T. Chamberlin)

Another case of piracy is shown on Plate IV. North and South lakes formerly drained westward past the site of

Haines Falls to Schoharie Creek, the present head of which is in the extreme northwest corner of the map. Kaaterskill Creek, which had a much higher gradient, worked headward and captured the headwaters of the westward-flowing stream, diverting to itself the drainage from North and South lakes. Schoharie Creek was thus beheaded.

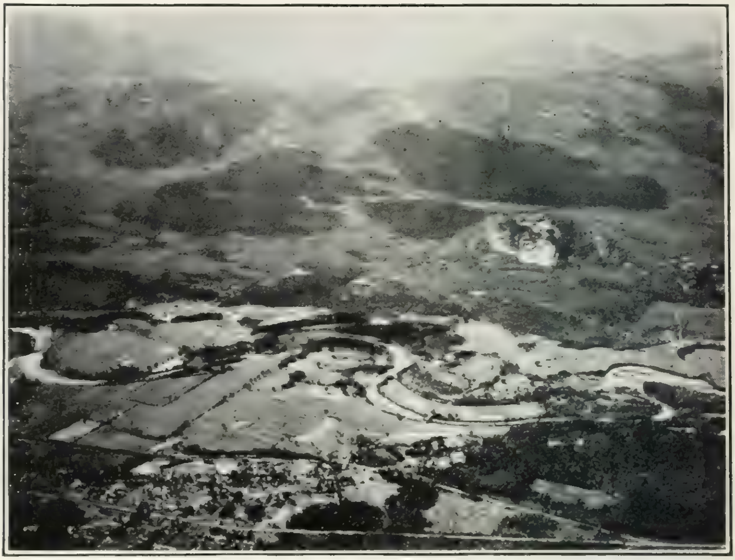
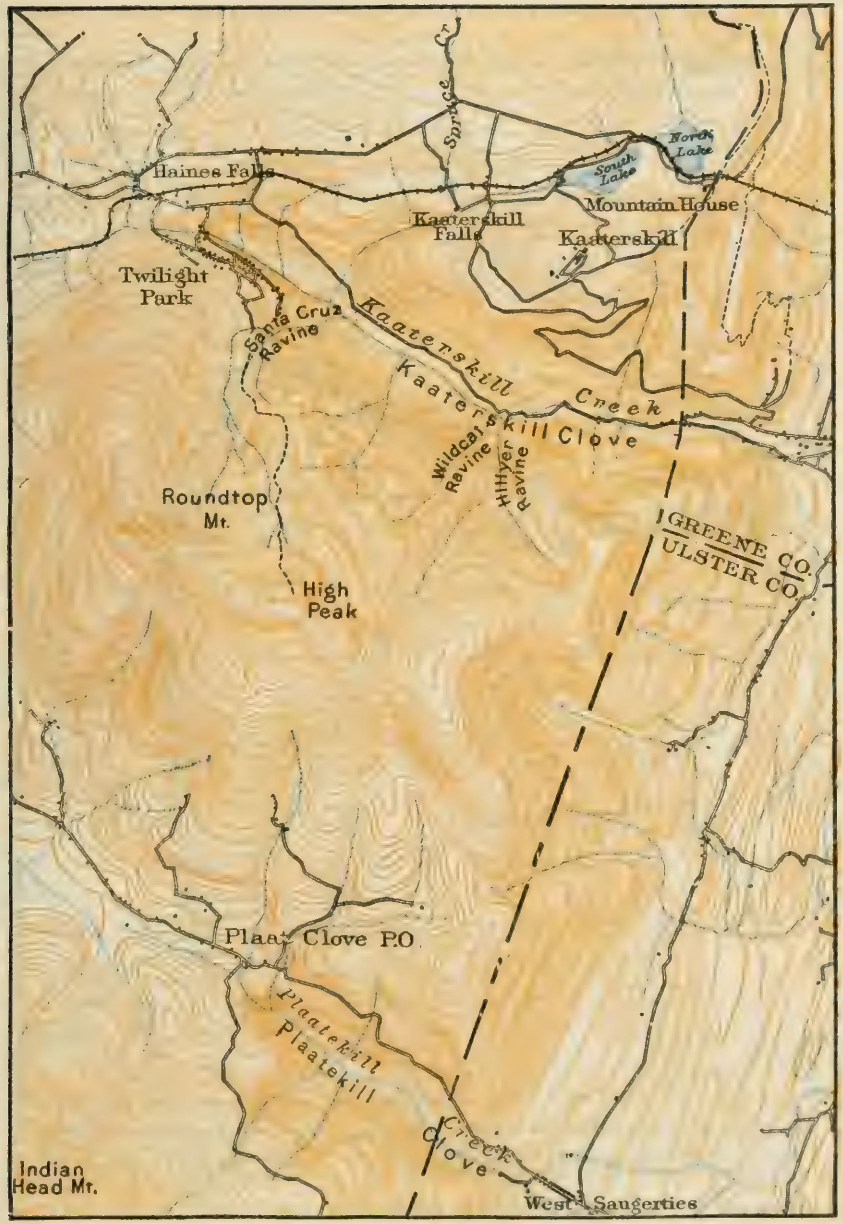


Fig. 87. — Mature stream valley. Soco River, N. H. (U. S. Army Air Serv.)

Plaatekill Creek, near the south limit of the map, appears to have beheaded the creek flowing west and northwest, similarly diverting its headwaters, and if it should continue unmolested in its piratical career it would capture two more tributaries of the same creek.

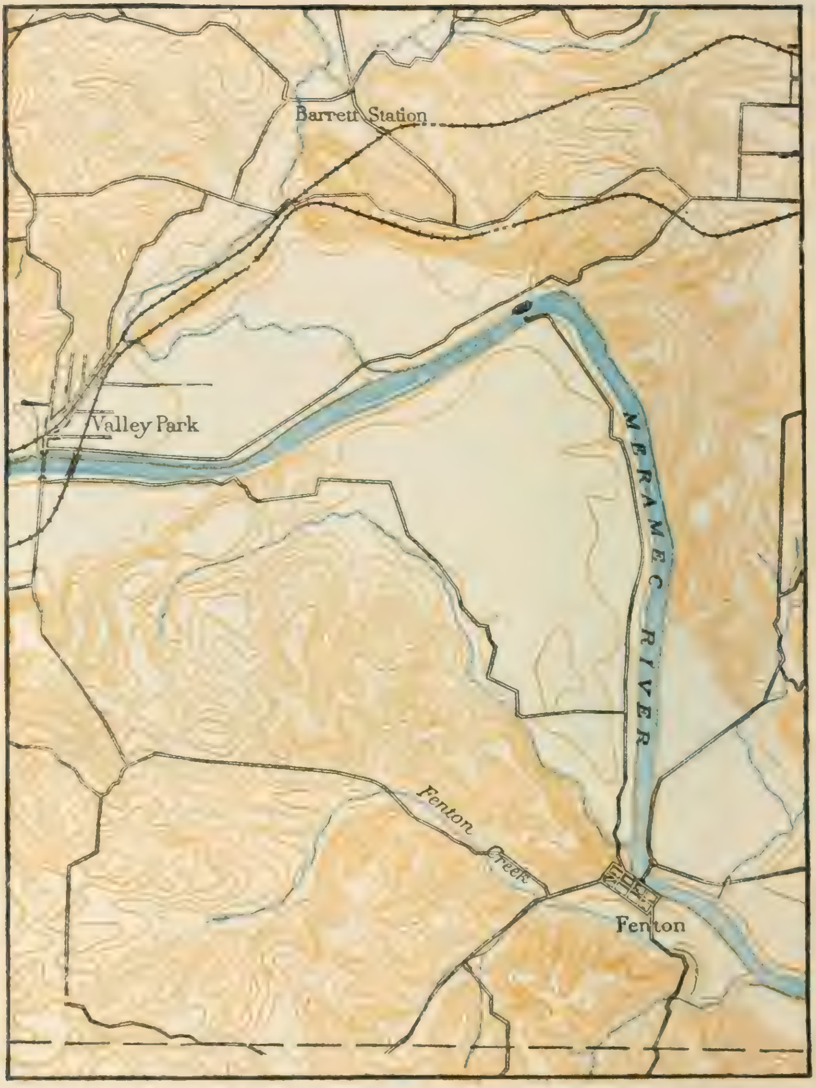
STAGES IN THE HISTORY OF A VALLEY

A valley passes through three more or less distinct stages during its growth: (1) While it is V-shaped in cross-section, and has a narrow bottom of steep gradient, with


Spruce Cr.
North Lake
South Lake
Haines Falls
Mountain House
Kaaterskill Falls
Kaaterskill
Twilight Park
Santa Cruz Ravine
Kaaterskill Creek
Kaaterskill Clove
Wildcat Ravine
Hillyer Ravine
Roundtop Mt.
GREENE CO.
ULSTER CO.
High Peak
Plaat Clove P.O
Plaatekill
Plaatekill
Creek
Clove
Indian Head Mt.
West Saugerties

Map illustrating piracy. The drainage from North Lake and South Lake was once westward to the westward flowing creek southwest of Haines Falls. The head of a tributary of Kaaterskill Creek worked back, capturing, near Kaaterskill Falls, the creek which formerly flowed westward, and diverted it southward to Kaaterskill Creek. Scale, about 1 mile per inch.

PLATE V


Barrett Station
Valley Park
MERAMEC RIVER
Fenton Creek
Fenton

A stream widening its valley by lateral planation. Scale, about 1 mile per inch. (U. S. Geol. Surv.)

few and short tributaries, it is *youthful* (Figs. 85 and 86). During this stage the most notable growth is in depth and in length. (2) When down-cutting has brought the bottom of the valley so low that the stream is no longer able to cut downward effectively, in other words is *at grade,* widening becomes dominant, the stream develops a valley flat, and the valley passes into *maturity* (Fig. 87). (3) When erosion has progressed so far that the river is slow and widely meandering, the flat wide, and the valley walls low and inconspicuous, the valley is *old.*

A CYCLE OF EROSION

The foregoing discussion should have made it clear that the topography of a region undergoing erosion changes greatly with the passage of time. The first effect of erosion by running water is to roughen the surface by cutting valleys and leaving ridges and hills. The final effect is to make it smooth again by cutting the ridges and hills down to the level of the valley bottoms.

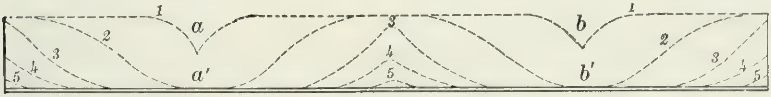


Fig. 88. — Diagram to illustrate the stages in an erosion-cycle, *a* and *b* are valleys in the youthful surface 1-1. When the valleys have grown to *a′* and *b′* the surface is mature. At stage 4-4 the topography of the region is old. At stage represented by 5-5 the region is a peneplain.

Base-level, Peneplain, Grade. A base-level of erosion has already been defined; but the mode of its development may now be illustrated in the light of the preceding discussion. Imagine a land surface affected by a series of parallel young valleys without tributaries (Fig. 88). Between them there is a series of upland plains or plateaus. The profile of the surface between two adjacent valleys is represented in section by the uppermost line in Figure 88. As the valleys are widened from 1 and 1 to 2 and 2, the intervening uplands are correspondingly narrowed. When the valleys have attained the form represented by 3 and 3, the intervening uplands have been narrowed to ridges, and the valley flats have become wide. With continued erosion the ridges will be lowered still more, and in time the land surface will approach a plain. In this condition it is known as a *peneplain*. The ridges will be obliterated finally, and a *base-level* plain will result. The time involved in the reduction of a land area to a base-level is a *cycle of erosion*.

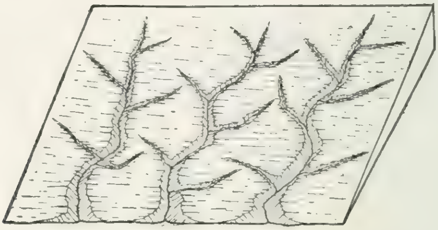


Fig. 89. — Diagram showing tributaries in an early stage of development.

Tributaries are almost sure to develop along each main valley and their heads to work back across the uplands between the main valleys, dissecting them into secondary ridges (Fig. 89). Tributaries will develop on the tributaries, and these tertiary valleys dissect the secondary ridges into ridges of a lower order. This process of tributary development goes on until drainage lines of the fourth, fifth, and sixth, and higher orders are formed (Fig. 90). Since the

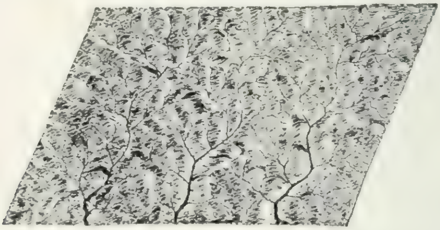


Fig. 90. — Diagrammatic representation of a surface much dissected by the development of numerous tributaries.

process of valley development under such circumstances is also the process of ridge dissection, a stage is presently reached in which the ridges are cut into sections so short that they cease to be ridges, and become hills instead. Even then the processes of erosion do not stop, for rain-water falling on the hills washes the loose material from their surfaces and starts it on its seaward journey. Thus the "everlasting hills" are lowered, and, given time enough, will be carried to the sea.

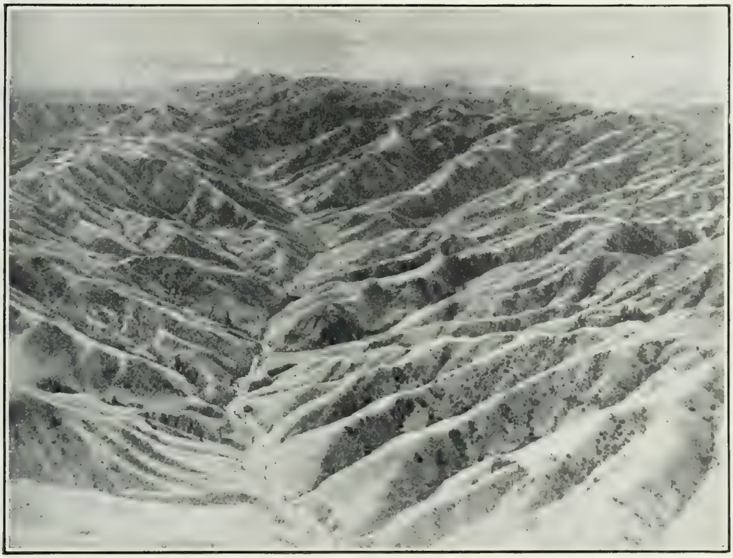


Fig. 91. — Thoroughly dissected mountains. Arrow Rock Dam, Idaho. (U. S. Army Air Serv.)

The base-leveled surface is not absolutely flat. The area reduced by each stream will have a slight slope downstream, and from its sides toward its axis. The low divides between streams flowing in the same direction may, however, disappear altogether, for when valleys have reached their limits in depth, their streams do not cease to cut laterally. Meandering in their flat-bottomed valleys, they often reach and undercut their divides (Pl. V, and Fig. 92). By *lateral planation*, the divides between streams may be entirely eaten away.

The ocean may be looked upon as a barrier which in a general way limits the down-cutting of running water. Other barriers, such as lakes and the outcrops of hard rock in stream beds, have a comparable, though more temporary, effect on the development of valley plains above themselves. Plains thus developed have been called *temporary base-levels.*

Stages in a Cycle of Erosion. (1) When considerable portions of a region are still unaffected by valleys, the region is in *topographic youth.* (2) When valleys have dissected the original upland surface and reduced the region predominantly to slopes, it is in *maturity.* And (3) when valley flats have grown so wide as to dominate the region, it is in *old age.* In youth running water has just begun its task of reducing the region to a base-level; in maturity the work is about half done; while in old age most of the work has been accomplished. The *topographic age* of a region is, therefore, *the ratio between the amount of work done and that still to be done* in reducing the region to a base-level. Topographic age is estimated without regard to tons or cubic miles of rock involved. A region in the interior of a continent may still be youthful though having suffered the

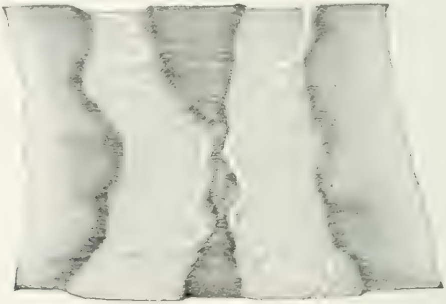


Fig. 92. — Diagram showing streams in adjacent valleys, undercutting the divide between them. They may, in time, destroy the divide by lateral planation.

PLATE VI

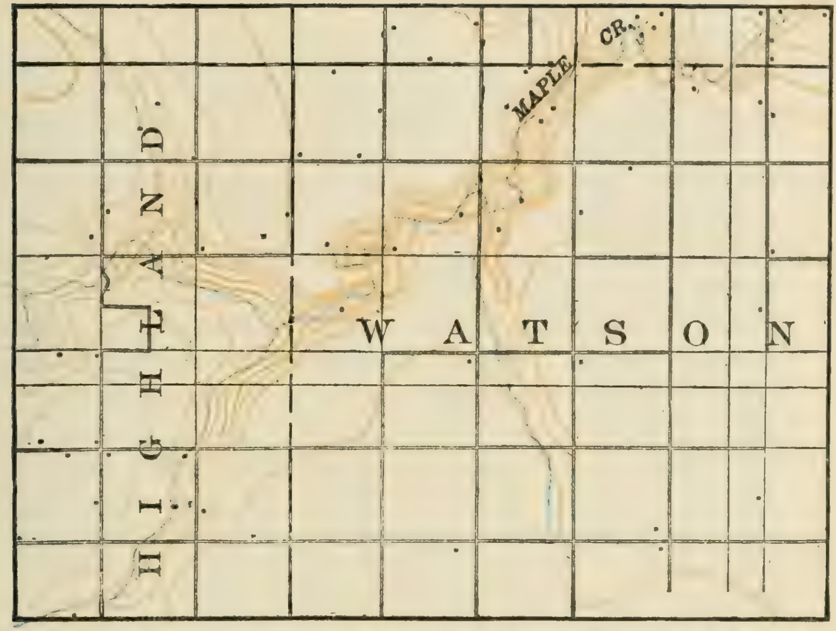

MAPLE CR.
HIGHLAND
WATSON

FIG. 1.—Youthful topography in a region of slight relief. The valley of Maple Creek is narrow, and much of the area is unaffected by erosion. Scale, about 2 miles per inch.

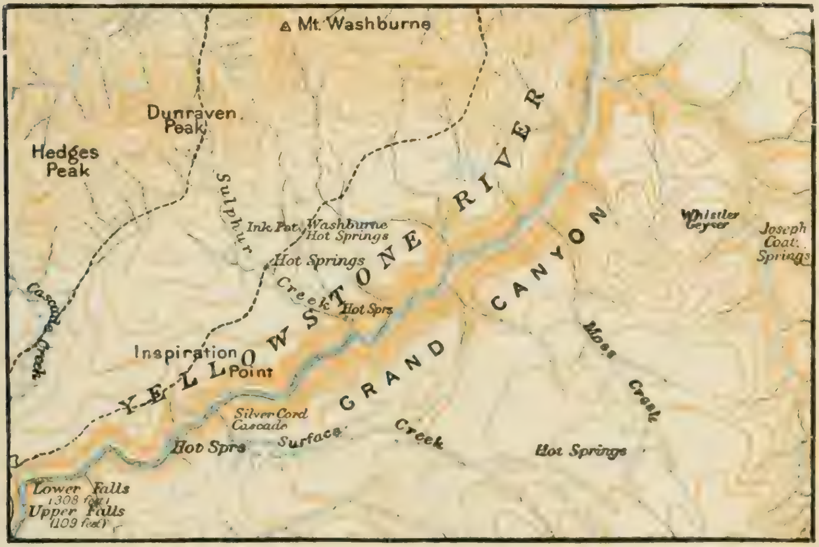

Mt. Washburne
Dunraven Peak
Hedges Peak
Sulphur Creek
Ink Pot
Washburne Hot Springs
Hot Springs
Hot Sprs
YELLOWSTONE RIVER
GRAND CANYON
Whistler Geyser
Joseph Coat Springs
Cascade Creek
Inspiration Point
Moss Creek
Silver Cord Cascade
Hot Sprs
Surface Creek
Hot Springs
Lower Falls (308 feet)
Upper Falls (109 feet)

FIG. 2.—A young valley in a region of great relief. Scale, about 2 miles per inch.

removal of many times the amount of rock material that has been eroded from a low-lying coastal region in old age. An erosion cycle may take millions of years, or it may, in miniature, be completed in the course of a single shower.

All of the valleys in a given region do not necessarily belong to the same stage of development as that of the region as a whole. Thus young valleys, if far from the main drainage lines, are to be expected even in old regions, while, on the other hand, mature valleys are not uncommon in young regions of slight elevation above a lake, or the sea.

The Erosion Cycle in an Arid Region. A radically different cycle is presented during the erosion of an arid region of original high relief. While, from arid regions, little or no water flows away as run-off, there is nevertheless a certain amount of erosion by running water *within* the area. For, even in arid regions, it does rain occasionally and sometimes very hard for short periods of time. Because the land is unprotected by vegetation, considerable erosion is accomplished during these sporadic downpours. During *youth* the inherited original roughness of topography of such a region is most manifest. As the cycle progresses, the uplands are worn down and the detritus deposited about the bases, coalescing in *maturity* to produce a continuous *piedmont alluvial plain.* The hills might be thought of as "islands" protruding above this "sea" of alluvium collecting about their bases. And finally, when the degraded uplands and the aggraded alluvial flats have come close to a common level, erosion by running water becomes very slow, and the region is said to be *old.* (It is to be noted that wind erosion, with the removal of quantities of material, is relatively very important in arid regions.)

GENERAL CHARACTERISTICS OF TOPOGRAPHIES DEVELOPED BY RIVER EROSION

With the characteristics of river valleys and the methods by which they grow clearly in mind, it is easy to say whether or not rivers have been the chief agents in the development of a given topography. River valleys are distinguished from other depressions on land surfaces (1) by their linear form, (2) by the fact that normal headward growth of tributaries develops a branching or *dendritic* drainage pattern in regions of massive or horizontally bedded rocks, and (3), leaving out of consideration the relatively insignificant inequalities in stream channels, by the fact that any point in the bottom is lower than any other point farther upstream in the same valley, and higher than any point farther downstream. The third point might be otherwise stated by saying that every valley excavated by erosion leads to a lower valley, to the sea, or to an inland basin. Streams which dry up, or otherwise disappear as they flow, constitute partial exceptions. If, therefore, the depressions on a land surface are linear, lead to other and deeper valleys, and finally to an inland basin or to the sea, and if the elevations between these valleys are such as might have been left by the excavation of the valleys, it is generally clear that running water has been the chief factor in the development of the topography. If, on the other hand, a surface is characterized by topographic features which streams cannot develop, such as enclosed depressions, or hills and ridges whose arrangement is independent of drainage lines, agents other than rain and surface streams have been concerned in its development.

RATE OF DEGRADATION

Since lands are being constantly eroded, it is of interest to ascertain how fast they are being degraded, and hence, how long they would last if the process continued without interruption.

Recent estimates indicate that the rivers of the United States carry annually about 270,000,000 tons of material to the sea in solution, and about 513,000,000 tons in suspension. This would mean that for the United States the average erosion is one foot in 8000–9000 years. Since the average elevation of the North American continent is not far from 2000 feet, erosion at the present rate, if nothing occurred to counteract it, would reduce the continent to sea-level in 16,000,000 to 18,000,000 years. But, as has already been seen, this rate of erosion will become less and less as the land is lowered. Long before the continent reached base-level, the rate of degradation, so far as dependent on mechanical erosion, would become so slow that the time necessary to bring the continent to sea-level would be indefinitely prolonged. Furthermore, the land is likely to suffer uplift relative to sea-level, which would offset erosion.

Conditions Affecting the Rate of Erosion. The rate at which running water erodes the land depends on (1) the character of the rock over which it flows, (2) the volume of water, (3) its velocity, and (4) the amount and kind of load it carries.

1. The *rock* over which the water flows influences the rate of its erosion in several ways: (a) If the rock is *physically weak*, being poorly indurated and of fine particles easily carried away, it is rapidly eroded. (b) If the rock is *chemically unstable*, is soluble, or has soluble cement between the particles, it is readily eroded. (c) If the structure of the rock presents many joints, cracks, bedding planes, or other weaknesses, it is more rapidly eroded than if there are fewer planes of weakness.

2. The greater the *volume* of water, other things being equal, the greater the erosion. The volume of sheet wash depends on the amount of precipitation, while that of a stream is governed not only by the amount of rainfall, but also by the size of the area draining into the stream.

3. The *velocity* depends upon (a) the gradient, (b) the volume of water, together with (c) the shape of the channel, and (d) the amount of the load. The steeper the gradient, the greater the volume, the narrower and deeper the channel, and the less the load, the swifter the flow of the water.

4. The *load* must be sufficient in amount to furnish

effective tools for abrasion and yet not great enough to retard notably the rate of flow. Furthermore, if the load is composed of hard, sharp particles, the stream obviously will erode more effectively than if the load consists of soft, rounded particles.

SPECIAL FEATURES RESULTING FROM SPECIAL CONDITIONS OF EROSION

Many striking topographic and scenic features result from rain and river erosion. Some of them depend primarily on the conditions of erosion, such as climate, altitude, etc., while others depend largely on the structure and resistance of the rock.

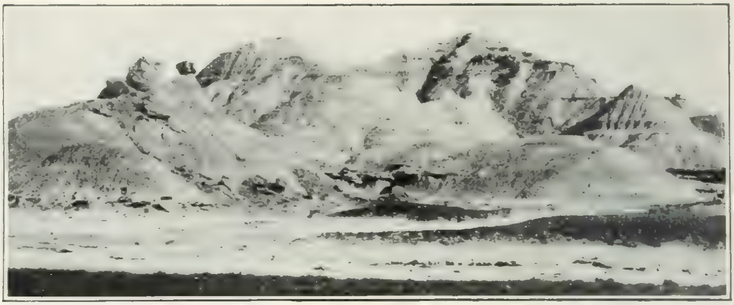
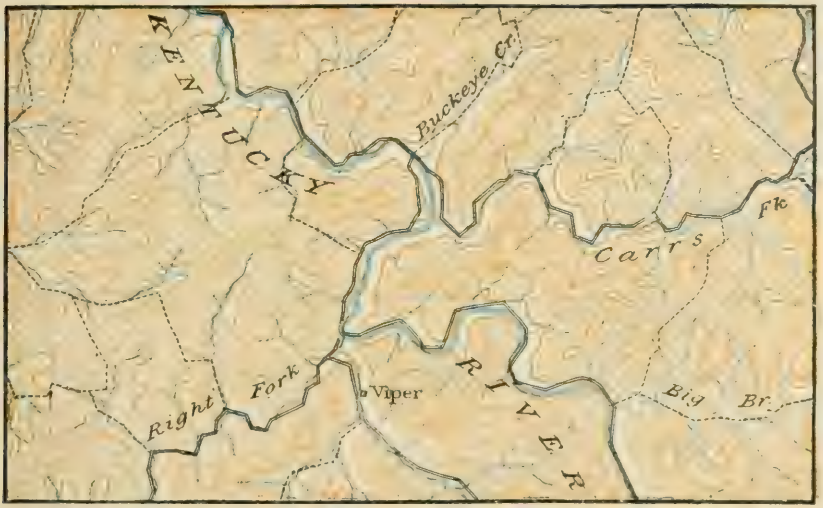


Fig. 93. — Bad lands of South Dakota. Oligocene formation. (Williston.)

Bad-land Topography. A type of topography consisting of a maze of gullies developed in early maturity in certain high regions where the rock is but slightly resistant, is termed *bad-land topography.* Some idea of bad-land topography is gained from Figure 93. Bad-land topography is found in many localities in the West, especially in western Nebraska and Wyoming, and in the western parts of the Dakotas, where the bed-rock is soft sandstone or shale, alternating with beds of clay. Climatic factors are also concerned in the development of bad-land topography. A semi-arid climate, where the rains, when they do occur,


KENTUCKY
RIVER
Buckeye Cr.
Carrs Fk
Big Br.
Right Fork
Viper

A region in a mature stage of erosion. Scale, about 2 miles per inch. (Kentucky, U. S. Geol. Surv.)

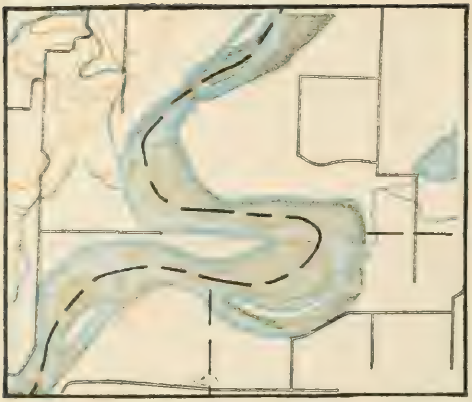


FIG. 1.—A meandering stream. The Missouri River. Scale, about 2 miles per inch. (Marshall, Mo., Sheet, U. S. Geol. Surv.)

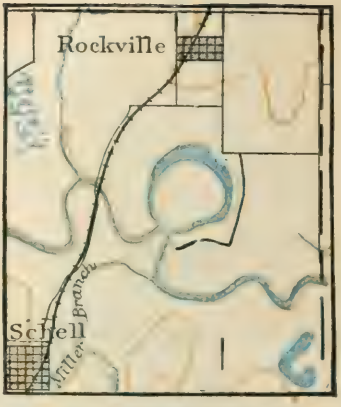

Rockville
Schell
Miller Branch

FIG. 2.—A stage in the development of a meander. Schell River. Scale, about 2 miles per inch. (Butler, Mo., Sheet, U. S. Geol. Surv.)

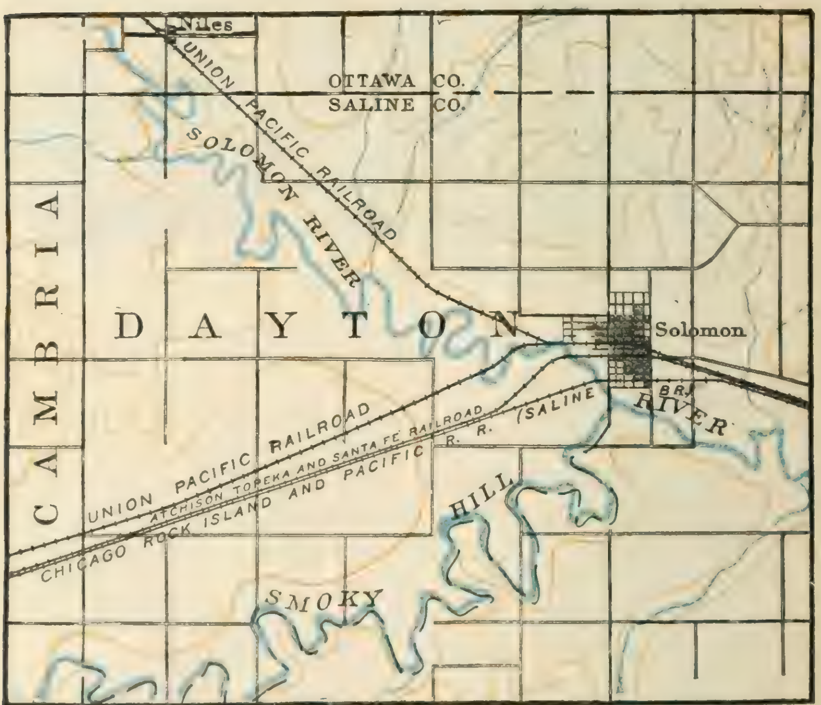

Niles
OTTAWA CO.
SALINE CO.
UNION PACIFIC RAILROAD
SOLOMON RIVER
CAMBRIA
DAYTON
Solomon
UNION PACIFIC RAILROAD
ATCHISON TOPEKA AND SANTA FE RAILROAD
CHICAGO ROCK ISLAND AND PACIFIC R. R.
(SALINE
BR.)
RIVER
HILL
SMOKY

FIG. 3.—A plain in old age. Scale, about 2 miles per inch. (Abilene, Kan. Sheet, U. S. Geol. Surv.)

are very heavy, seems to be the most favorable for its development.

Canyons. Various conditions influence the size and shape of valleys, especially in the early stage of their development. High altitude of land favors swift-flowing streams and the development of deep valleys. Such valleys will be narrow if the conditions which determine widening are absent or unfavorable. An arid climate favors the development of narrow valleys, if there is sufficient water to maintain a vigorous stream, because there is little slope wash. Narrow valleys with steep slopes will also be favored if the valley is cut in rock which is capable of standing with steep faces. Thus a stream may develop a narrow valley in firm rock where it would not do so in loose gravel. Aridity, high altitudes, and the proper sort of rock therefore favor the development of canyons, and many of the young valleys in the western part of the United States, where these conditions prevail, belong to this class.

While all canyons are valleys, most valleys are not canyons. In popular usage, the rule seems to be that if a valley is sufficiently deep, narrow, and steep-sided to be distinctly striking, it is called a canyon in regions where that term is used. Whether a valley is deep, narrow, and steep-sided enough to be striking, clearly depends on its surroundings and on the observer. The Colorado Canyon (Fig. 94) is the greatest canyon known, but only in its deepest section is it more than a mile deep, and where its depth approaches this figure it is 8, 10, or even 12 miles wide from rim to rim. Its width at the bottom is little more than the width of the stream; that is, a few hundred feet. Its cross-section throughout much of its course is therefore not in keeping with the conventional idea of a canyon. With a depth of one mile and a width of eight, the slope, if uniform, would have an angle of less than fifteen degrees. Such a valley is represented in Figure 95. As a matter of fact the slopes of a canyon are not uniform, but more like those in

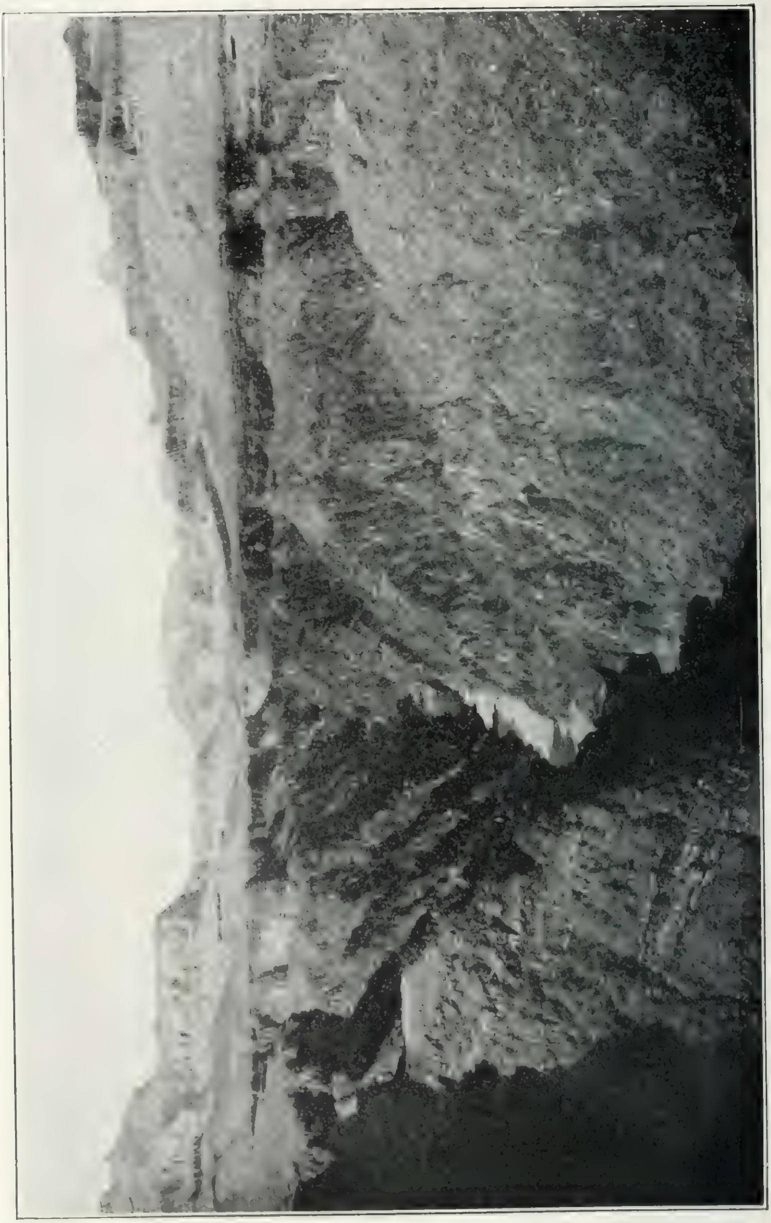


Fig. 94. — Grand Canyon of the Colorado. (Carkhuff.)

Figure 96. The step-like slopes are due to inequalities of hardness, and it is the first drop of more than 1000 feet below an observer on the rim of the canyon that gives the tremendously striking effect.

Like all valleys which are narrow relative to their depth, the Colorado Canyon, gigantic as it is, is a young valley, for it represents but a small part of the work which the stream must do to bring its drainage basin (the Colorado Plateau) to base-level.

Fig. 95. — Diagram showing the proportions of a valley the width of which is eight times the depth. These are approximately the proportions of the Colorado Canyon.

While aridity and high altitude are conditions which favor the development of canyons, as shown by the fact that most canyons are in high and dry regions, they are not indispensable. Niagara River has a canyon below its falls (Pl. IX), and the surrounding region is neither high nor arid. The narrow part of the valley is so young that side erosion has not yet widened the valley, or lowered its angle of slope, to such an extent as to destroy its canyon-like character. This canyon is often called a *gorge*, a term frequently applied to small valleys of the canyon type. (Fig. 97).

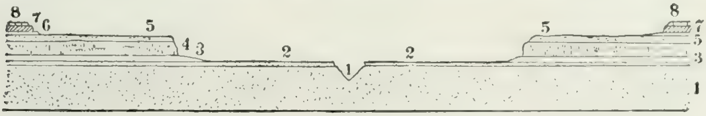


Fig. 96. — Cross-section of the Colorado Canyon. (After Gilbert and Brigham.)

EFFECTS OF UNEQUAL HARDNESS

Irregularities of hardness give rise to many peculiarities of topography and to many scenic features. To this

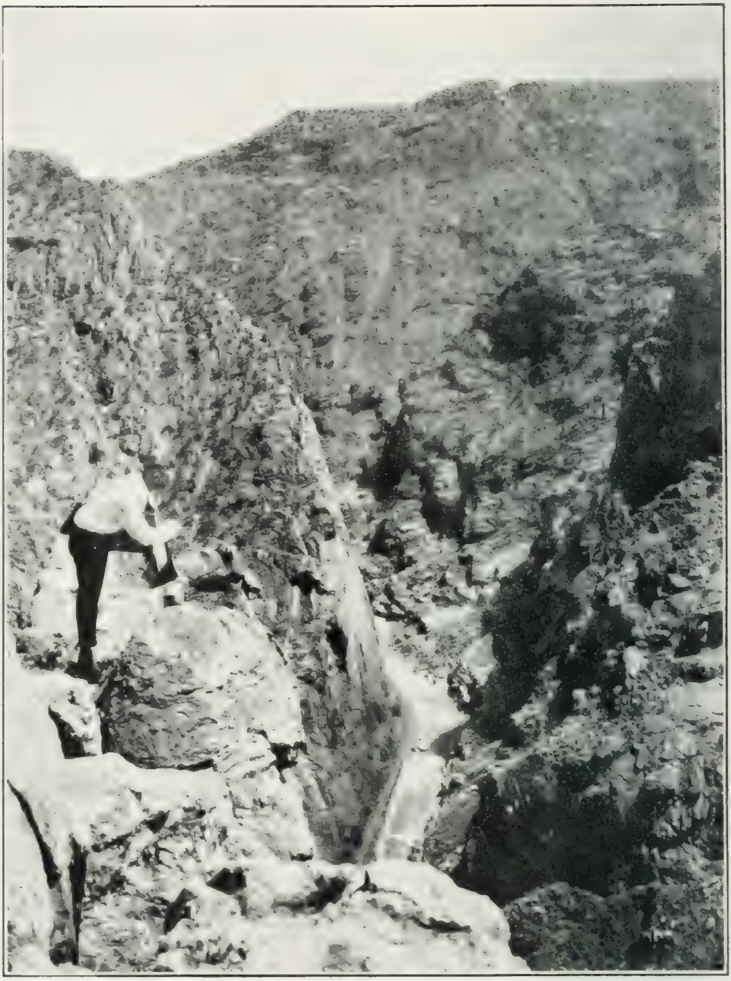


Fig. 97. — Royal Gorge of Arkansas River, Colo.
(Courtesy U. S. Forestry Serv.)

category belong many rapids, falls, narrows, terraces, and many striking hills and ridges.

Falls and Rapids. Where the flow of a stream is notably quickened, it makes rapids; where the descent is vertical, or nearly so, it is a waterfall. Falls and rapids are most commonly developed where streams pass from more resistant to less resistant rock. The faster erosion of the latter gives rise to rapids. At first the rapids are slight (*a*, Fig. 98), but they become more considerable (*b*) as time and erosion go on. When the bed of the rapids becomes sufficiently steep, the rapids become *falls*. When the water *falls* rather then *flows* over the rock surface below the hard layer, erosion assumes a new phase. The hard layer is then undermined, and the undermining causes the falls to recede. This phase of erosion is sometimes called *sapping* (Fig. 99).

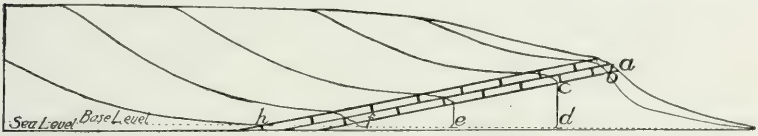


Fig. 98. — Diagram illustrating the development of a fall where the hard layer dips upstream.

If the hard layer which occasions a fall dips upstream (Fig. 98), its outcrop in the stream's bed becomes lower as the fall recedes (*e*). When it has become so low that the water passing over it no longer reacts effectively against the less resistant material beneath (*f*), sapping ceases, and the fall is then transformed again into rapids. The history of rapids which succeed falls is the reverse of the history which preceded. The later rapids are steepest at

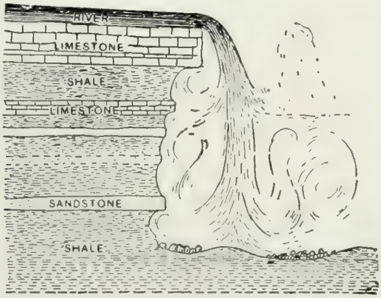


Fig. 99. — Diagram illustrating the conditions at Niagara Falls. (Gilbert.)

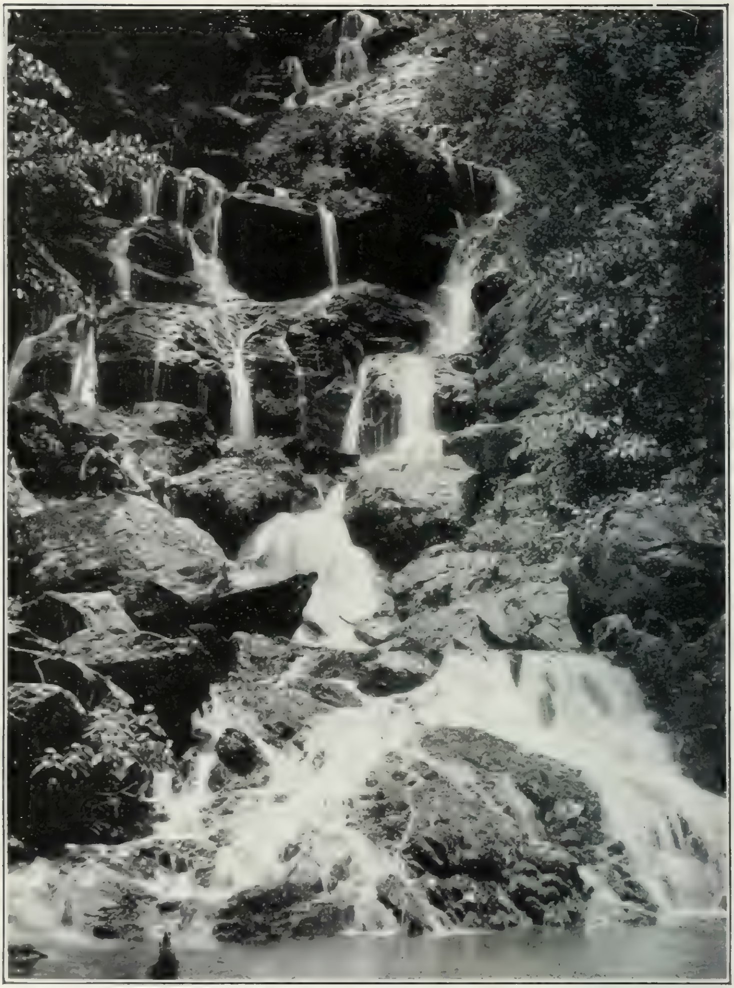


Fig. 100. — Cascade. Pisgah National Forest, N. C.
(Courtesy U. S. For. Serv.)

the beginning of their history, the earlier at their end. Stated in other terms rapids are steepest when nearest falls in time. Slight differences in the resistance of successive layers of rock often occasion successive falls or rapids.

If the layers of unequal hardness in a stream's bed are vertical and cross the course of the stream at right angles, rapids and perhaps falls will develop. Falls developed under these conditions will not recede.

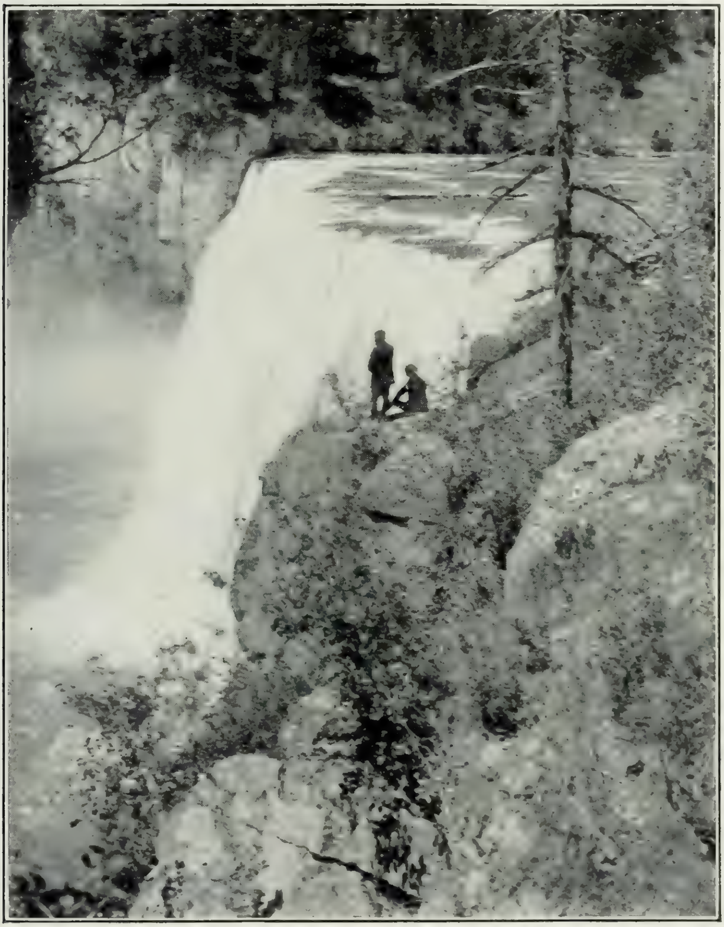


Fig. 101. — Big Falls, Snake River. Taighee Forest, Id. (Courtesy U. S. For. Serv.)

The inequality of resistance in the rock which occasions a waterfall may be original or secondary. In the case of Niagara Falls (Fig. 99) relatively resistant limestone overlies relatively weak shale. At the Falls of St. Anthony (Minneapolis) limestone overlies friable sandstone. The Falls of the Yellowstone and the Shoshone Falls of the Snake River (Idaho) are in igneous rock. In the former case the unequal resistance is occasioned by unequal decay of the rock, due perhaps to the rise of hot vapors which have decomposed the rock along the lines of their ascent; in the latter, a more resistant sort of igneous rock overlies a less resistant one.

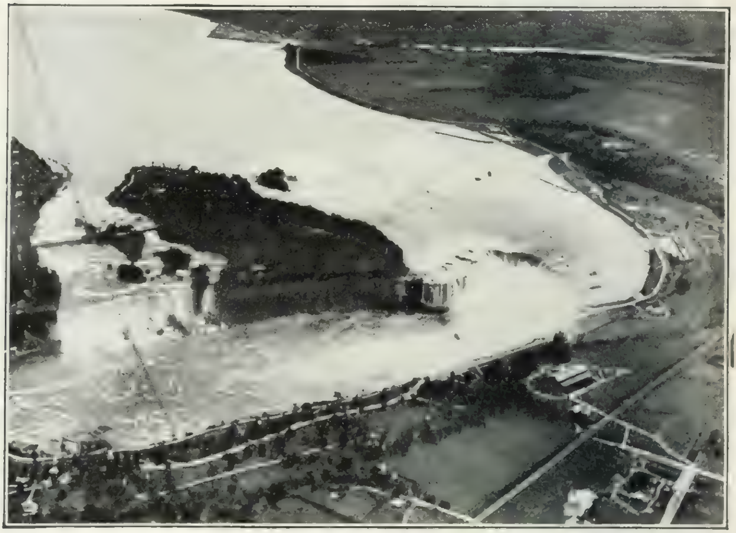
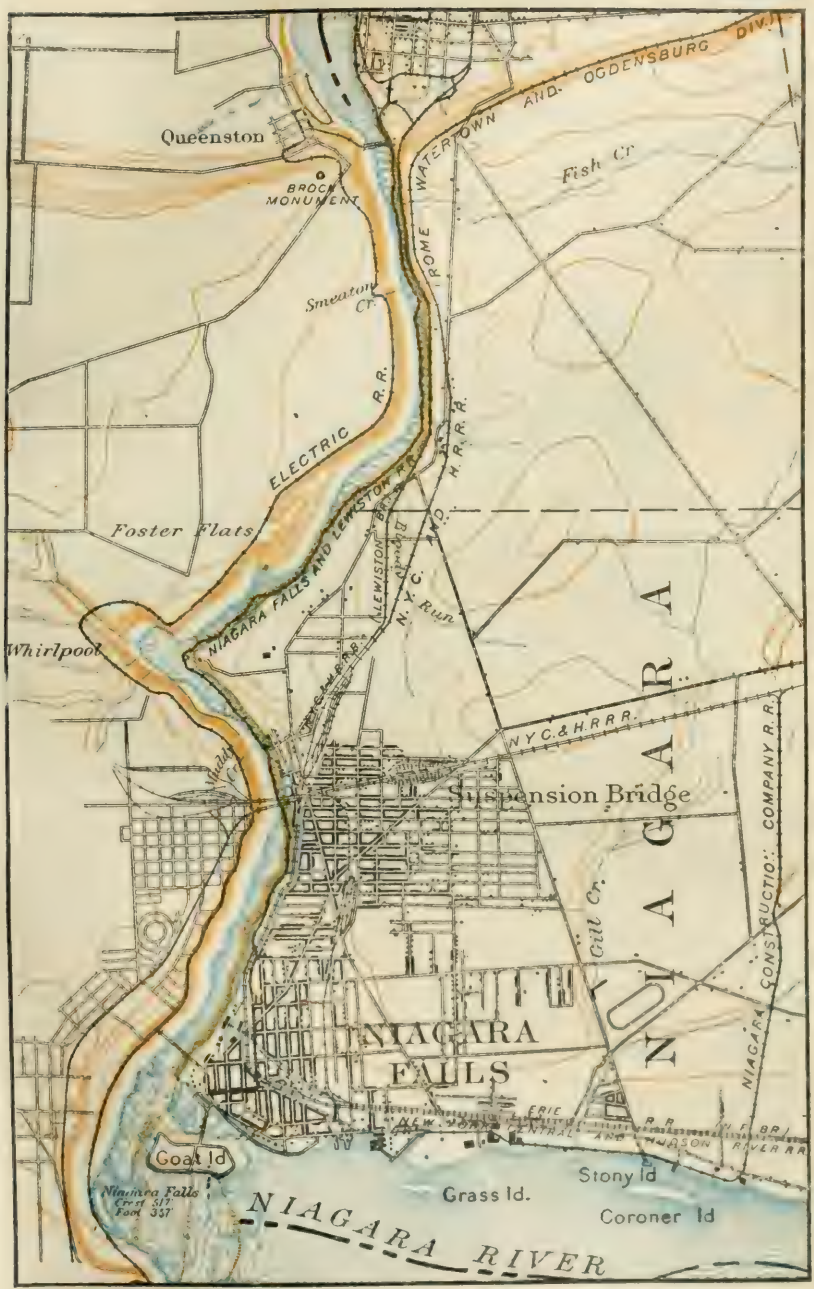


Fig. 102. — Niagara Falls from the air. (Kadel and Herbert.)

One waterfall often breeds another. Thus where a fall recedes beyond the mouth of a tributary stream, the latter is also caused to fall. The Falls of Minnehaha Creek, tributary to the Mississippi near Minneapolis, are an illustration. Once in existence, the fall of a tributary follows the same history as that of the main stream.

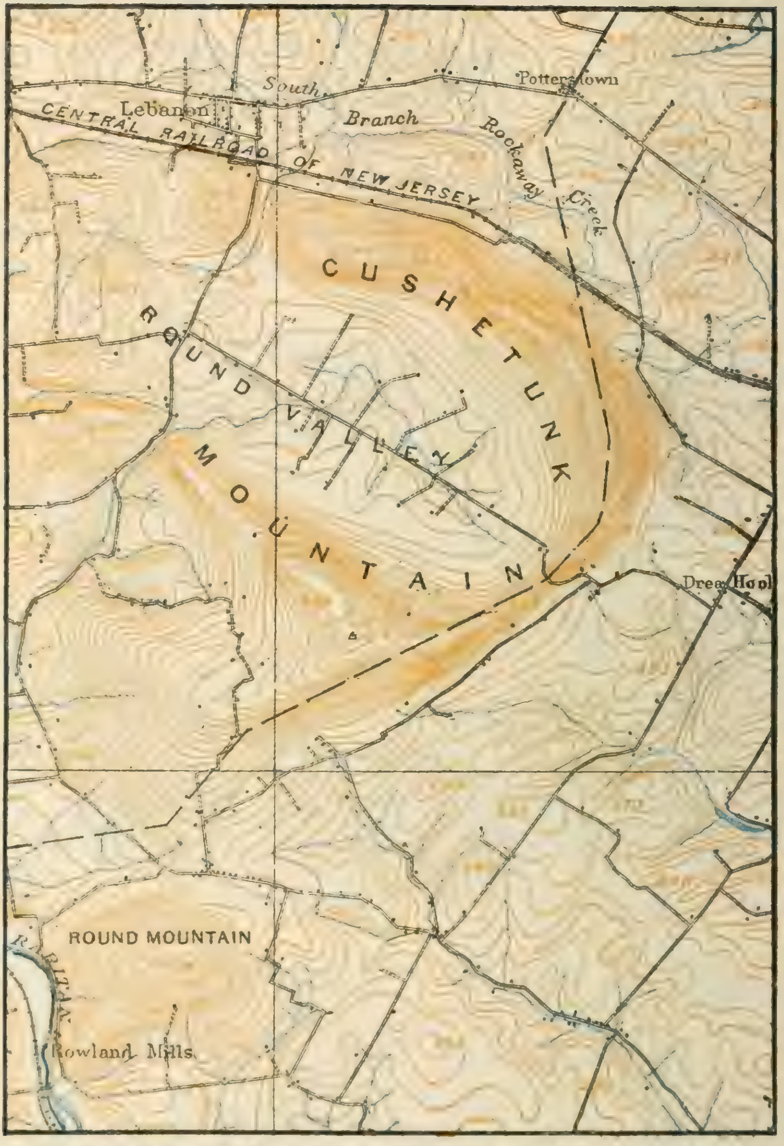
Niagara Falls (Plate IX) are remarkable both because of the large volume of water and the great descent which is between 160 and 170 feet. These falls are divided into two parts, separated by Goat Island, the Horseshoe Fall (Fig. 102)

PLATE IX


Queenston
BROCK MONUMENT
WATERTOWN AND OGDENSBURG DIV.
ROME
Fish Cr
Smeaton Cr.
ELECTRIC R.R.
N.Y.C. AND H.R.R.R.
NIAGARA FALLS AND LEWISTON R.R.
LEWISTON
Foster Flats
Whirlpool
Bloody Run
N.Y.C. & H.R.R.R.
Suspension Bridge
NIAGARA
Gill Cr.
NIAGARA CONSTRUCTION COMPANY R.R.
NIAGARA FALLS
ERIE R.R.
NEW YORK CENTRAL AND HUDSON RIVER R.R.
Goat Id
Niagara Falls
Crest 517
Foot 337
Stony Id
Grass Id.
Coroner Id
NIAGARA RIVER

The Niagara Gorge. Scale, about 1 mile per inch. (Niagara Falls, N. Y., Sheet, U. S. Geol. Surv.)

PLATE X


Potterstown
South
Lebanon
Branch
CENTRAL RAILROAD OF NEW JERSEY
Rockaway Creek
CUSHETUNK
ROUND VALLEY
MOUNTAIN
Dreahook
ROUND MOUNTAIN
RARITAN
Rowland Mills

Cushetunk and Round Mountains, New Jersey; examples of isolated mountains left by the removal of less resistant surroundings. Scale, about 1 mile per inch. (High Bridge Sheet, U. S. Geol. Surv.)

on the south, and the American Fall on the north. Between 1842 and 1905, the Horseshoe Fall receded about five feet a year, while between 1827 and 1905, the American Fall receded less than three inches a year.

Rapids and falls sometimes occasion the development of pot-holes in the rock at the foot of the falls, a peculiar rather than important feature of erosion. The holes are excavated in part by the falling and eddying of silt-charged water, but chiefly by stones whirled in the eddies. Pot-holes which are not now in immediate association with rapids or falls point to their former existence in these places.

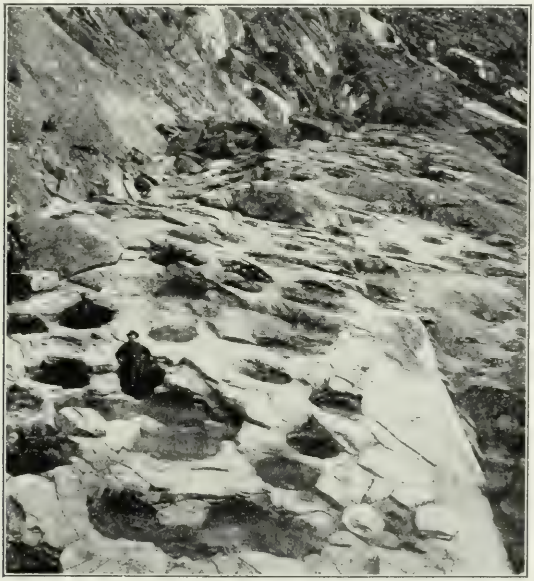


Fig. 103. — Pot-holes in granite. Upper Tuolumne River, Calif.

Rock Terraces. Where a hard layer outcrops in the side of a valley above its bottom, the side slopes of the valley become gentle just above the hard layer, and steep, or even vertical, at and below its outcrop, as illustrated by Figure 104. The hard layer then stands out as a rock terrace on one or both sides of the valley.

Narrows. If a stream crosses vertical or highly inclined strata of unequal resistance, its valley is usually constricted at the crossing of the resistant layers. If such a constriction is notable it is called a *narrows*, or sometimes a *water-gap* (Fig. 105). The Appalachian Mountains afford numerous examples. The narrows develop because the processes which widen the valleys are less effective on the resistant layer than on the less resistant ones. Narrows also arise in other ways.

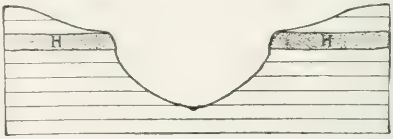


Fig. 104. — Rock terraces, due to resistant layers of rock.

Narrows are more conspicuous in certain stages of the erosion cycle than in others. While a valley is still so young as to be narrow at all points, there can be no pronounced "narrows"; but later, when the valley is otherwise wide, narrows are more pronounced. From what has preceded it is clear that rapids or falls are likely to occur at narrows, especially in the early part of their history.

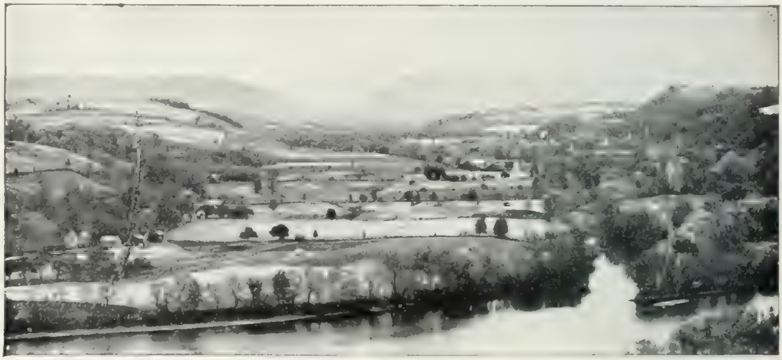


Fig. 105. — The Kittatinny Mountain and Delaware Water-Gap from Manunka Chunk. (N. J. Geol. Surv.)

Other Effects on Topography. Inequalities in the hardness of rock develop certain peculiarities of topography outside of valleys. The less resistant portions of a land area more or less distant from streams are worn down more readily than those which are more resistant. If great areas of high land be capped with hard rock, they are likely to remain as plateaus, or *mesas*, after surrounding areas of less resistance are brought low. If the hard capping remains over a small area instead of a large one, the elevation is a *butte*, a hill or a mountain, instead of a plateau (Fig. 106). Many buttes and small mesas are but remnants of former plateaus (Fig. 107). A feature of buttes and mesas capped by hard

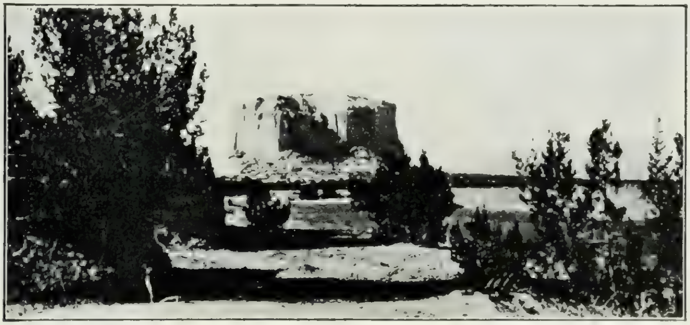


Fig. 106. — The Enchanted Mesa. A striking butte in New Mexico. The name mesa is not commonly applied to elevations of such small summit area. (R. T. Chamberlin.)

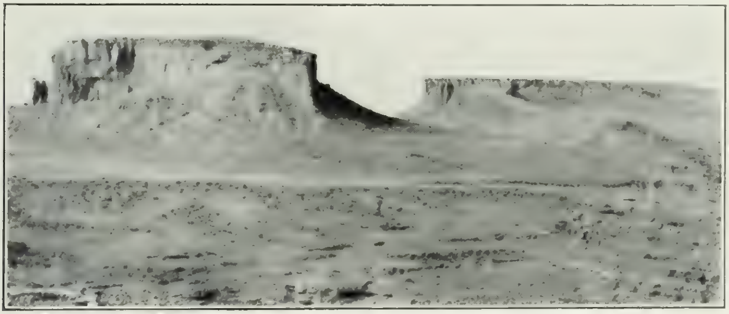


Fig. 107. — Frewen's Castle, an isolated butte. (Wegeman, U. S. Geol. Surv.)

rock is the steep bounding slopes or cliffs corresponding to the edges of the hard beds.

If the rock of a region is stratified and the layers tilted, the removal of the softer beds leaves the harder ones projecting above the general level in the form of ridges or "hog-backs" (Fig. 108). Dikes of igneous rock, harder than the beds which they intersect, likewise become ridges after the degradation of their surroundings. The plugs of old volcanic vents and other igneous intrusions of limited area often constitute conspicuous hills or mountains after erosion has removed their less resistant surroundings (Fig. 109). Round Valley, shown on Plate X, is an example.

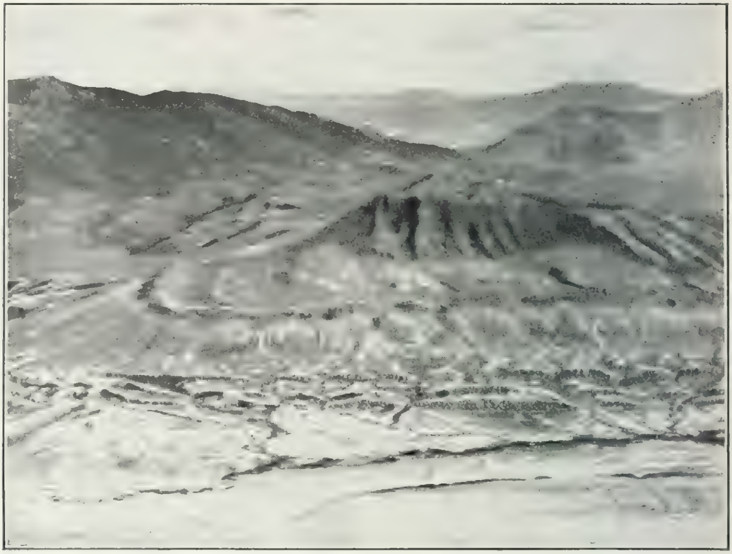


Fig. 108. — Hogbacks flanking the western scarp of Medicine Bow Range, Wyo. (Courtesy U. S. Army Air Serv.)

Ridges and hills resulting from the unequal degradation of unequally resistant strata are not equally prominent at all stages in an erosion cycle. In early youth the material surrounding the hard bodies of rock has not been removed; in early maturity considerable portions of their surroundings still remain about them; but in late maturity or early old age the outcropping masses of hard rock have been more perfectly isolated, and are most conspicuous. Most of the even-crested ridges of the Appalachian Mountain system as well as many others which might be mentioned, became ridges in this way. In the final stages of an erosion cycle, the ridges of hard rock are themselves brought low. Isolated remnants of hard rock which remain distinctly above their surroundings in the late stages of an erosion cycle (Fig. 110) are known as *monadnocks*, the name being derived from Mount Monadnock, N. H., an elevation of this sort.

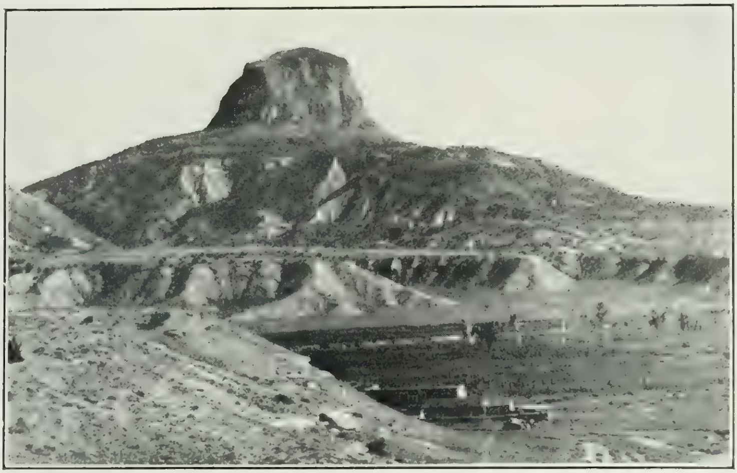


Fig. 109. — A monadnock. Canejon Butte, New Mex. Volcanic plug surrounded by weak sandstone and shale. (Lee, U. S. Geol. Surv.)

Adjustment of Streams to Rock Structure. Many valleys (gullies) are located at the outset without immediate regard to the hardness or softness of their beds. It is primarily the slopes about the heads of such gullies which determine their lines of growth, producing consequent streams, and once established, streams tend to hold their courses. Streams on weaker rock will deepen their valleys more rapidly than others, while those which flow across stronger and weaker rock alternately, will deepen their valleys more rapidly than those which run continuously on hard rock.

The streams on the weaker rocks, therefore, have an advantage over the others. Being deeper, their tributaries may be lengthened until their heads reach the other valleys, with the results shown in Figures 111-113. Even where several streams cross the same resistant bed, piracy is likely to take place among them, for some are sure to deepen their valleys faster than others, because of inequalities of volume, load, or hardness of bed. This is illustrated by Figures 114–116. An actual case is shown in Figures 83 and 84. Though piracy may take place where streams do not flow over rock of unequal resistance, it is much more common where they do, for unequal resistance puts one stream at a disadvantage compared with another.

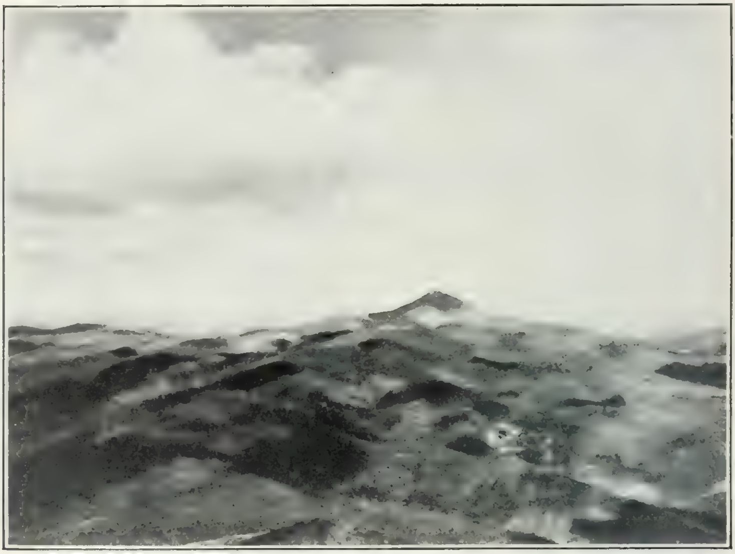
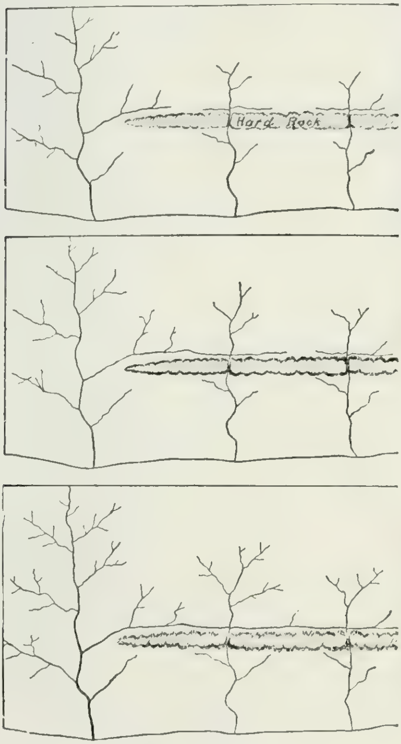
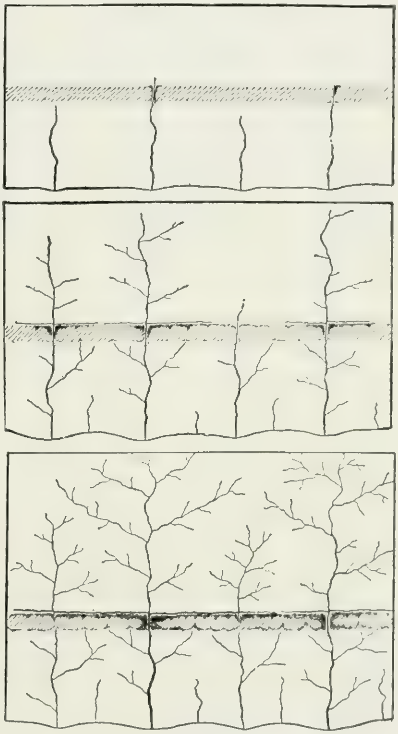


Fig. 110. — Peak of Conçeiçâo. Minas Geraes, Brazil. A monadnock of iron ore which owes its elevation to resistance greater than that of its surroundings. (R. T. Chamberlin.)

The changes in the courses of streams, by means of which they come to sustain definite and stable relations to the rock structure beneath, are known as processes of *adjustment.* Since streams and valleys adjust themselves to other conditions as well, this phase of adjustment may be called *structural adjustment.* Structural adjustment is not uncommon among rivers flowing over strata which are vertical or highly inclined, since in these positions the hard and soft strata are most likely to come to the surface in frequent alternation.



Figs. 111–113



Figs. 114–116

Figs. 111–113. — Diagrams illustrating piracy, where the stream which does not flow over rock of greater hardness captures those which do. Fig. 112 represents a further development of the drainage shown in Fig. 111, and Fig. 113 represents a still later stage.

Figs. 114–116. — Diagrams to illustrate piracy where the competing streams all cross a hard layer. The diagrams represent successive stages of development.

The processes of adjustment go on until, when completed, the streams are flowing as much as possible on the weaker beds, and as little as possible on the stronger. This amounts to the same thing as saying that the outcrops of the hard layers become the divides. In many cases an area is so located that there is no escape for its drainage except across resistant rock. In these cases the drainage is completely adjusted when as few streams as possible cross the resistant rock, and these by the shortest routes.

Adjustment has been carried to a high degree of perfection in most parts of the Appalachian Mountains. Here, as in all other mountains of similar structure, strata of unequal hardness were folded into ridges. The folds were then truncated by erosion, exposing the more and the less resistant beds (*H* and *S*, respectively, Fig. 117) in alternate belts along the flanks of the truncated folds (truncated at *ab* and *cd*). The streams, especially the lesser ones, now flow along the outcrops of the weaker beds much more extensively than elsewhere, and where they cross the hard layers it is usually at right angles. This is shown in Figure 118, where the trend of the outcrops is northeast-southwest.

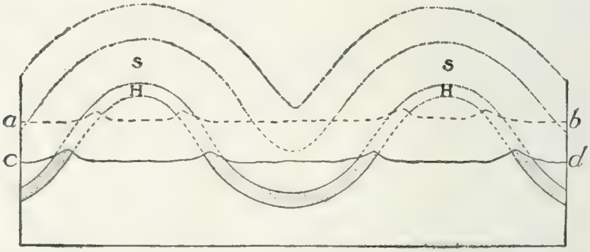


Fig. 117. — Diagram showing the outcrops of hard layers (shaded) on the flanks of truncated folds: *cd*, present surface; *ab*, an earlier erosion surface.

As base-level is approached, the outcrops of hard rock are brought low. When they have been reduced to the level of their surroundings, the streams may flow without regard to the resistance of the rock beneath, for the downward cutting has ceased.

It sometimes happens that rocks of unequal resistance are covered by beds of uniform hardness. A consequent

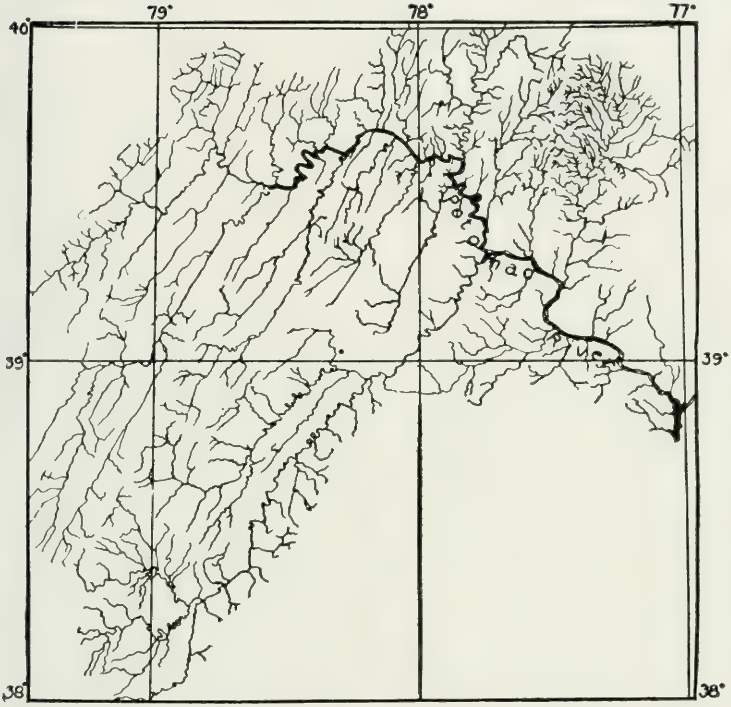

40°
79°
78°
77°
39°
39°
38°
38°
Potomac
River

Fig. 118. — Streams of the upper Potomac Basin, illustrating adjustment.

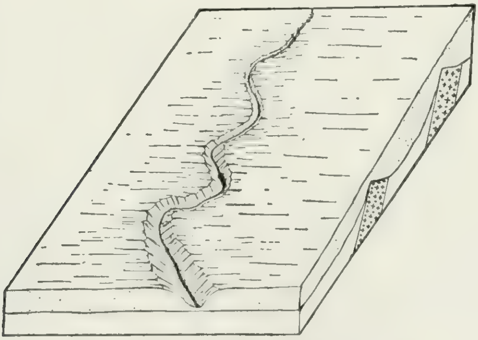


Fig. 119. — Diagram to illustrate superimposition. The consequent stream on the upper formation became superimposed on the underlying structures when the upper bed had been cut through.

stream developed on the latter may find itself out of structural adjustment when it has cut its channel down to the level of the heterogeneous beds below. Such a stream is said to be *superimposed* (Fig. 119) on the underlying structure. Structural adjustment is likely to follow in time.

Influence of Joints. The large, more or less vertical cracks in rock formations are known as joints. There are commonly two or more sets crossing each other at high angles. These joints allow the ingress of water, roots, etc., which help to weather and disrupt rocks. Their effect on erosion may often be seen along such a stream as flows in a rock-walled gorge. In such a situation, the outlines of the banks are in places angular, and in places crenate (Fig. 120), the re-entrants being located at the joints. By working into and widening these joints, running water sometimes isolates masses of rock to form islands.

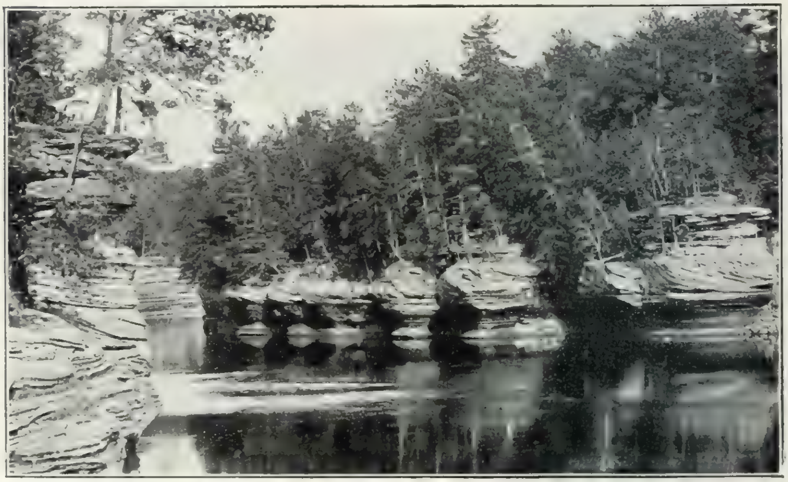


Fig. 120. – Figure showing crenate river bank, the re-entrants being determined by joints. Dells of the Wisconsin River, near Kilbourn, Wis. (Atwood.)

In a region free from mantle rock, or where the mantle rock is meager, joints often determine the courses of valleys by directing the courses of surface drainage. This is shown in many parts of the arid West. In regions where the rocks are notably faulted, (see p. 304), the courses of the streams are, in many cases, controlled by the fault planes.

The jointing of rocks often shows itself distinctly in the weathered faces of cliffs (Figs. 121 and 122), especially in arid and semi-arid regions, or where the slope is too steep for the accumulation of soil and rock-waste on its surface.

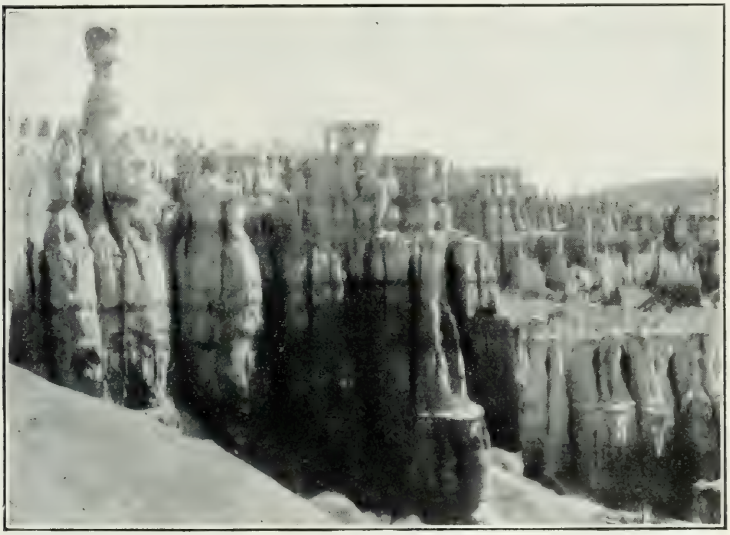


Fig. 121. — Bryce Canyon, Utah. (Courtesy U. S. Forestry Serv.)

Joints in rocks may lead to the development of natural bridges. If, above a waterfall, for example, there were an open joint in the bed of the stream (as in Fig. 123), some portion of the water would descend through it. After reaching a lower level it might find or make a passage through the rock to the river at the base of the fall. If even a little water took such a course, the flow would enlarge the passageway through the joint to the valley at the fall. This passageway might in time become large enough to accommodate all the water of the river. In this case, the entire fall would be transferred from the position which it previously occupied to the position of the enlarged joint. The fall would then recede. The underground channel between the old fall and the new would then be bridged by rock (Fig. 124). The natural bridge near Lexington, Va. (Fig. 125), almost two hundred feet above the stream which flows beneath it, is believed to have been developed in this way, but it is not to be understood that all natural bridges have had this history.

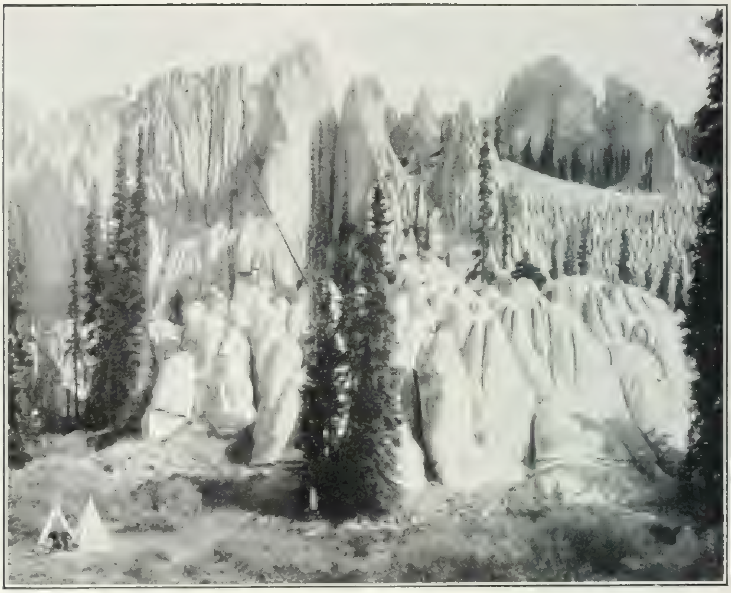


Fig. 122. — The "Ghosts." Wheeler Natural Monument, Rio Grande National Forest, Colo. (Courtesy U. S. For. Serv.)

Folds. The erosion of folded strata (*anticlines* or upfolds and *synclines* or downfolds) leads to the development of distinctive topographic features. So soon as an anticline begins to be lifted, it is, by reason of its position, subject to more rapid erosion than its surroundings. For the same reason, the upper portion of the fold is likely to be degraded more rapidly than its lower slopes, and must suffer more degradation before it is brought to base-level. Since most folds are composed of beds of unequal resistance, as the degradation proceeds and successive layers are worn from their upper portions, alternating hard and soft layers become exposed. As this proceeds, adjustment of the streams is likely to cause the water courses, and later the valley plains, to be located on the outcrops of the less resistant layers, leaving the stronger beds as ridges.

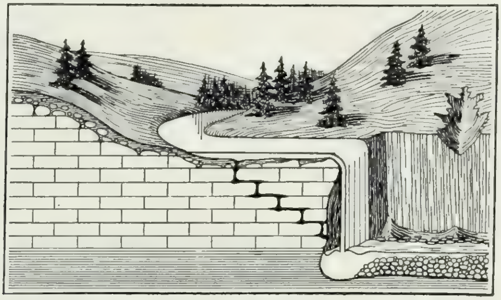


Fig. 123. — Diagram illustrating conditions for the formation of a natural bridge. (Krumbein.)

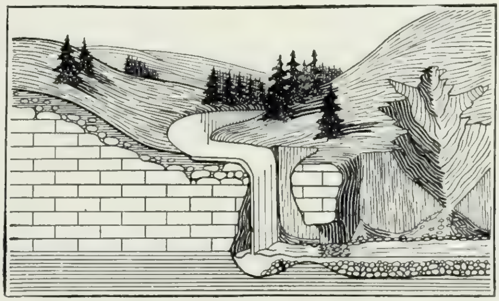


Fig. 124. — Diagram of same region as Fig. 123 after the natural bridge has been formed. (Krumbein.)

If the axis of an eroded anticline is horizontal, a given hard layer, the arch of which has been cut off will, after erosion, outcrop on both sides of the axis. When the topography has become mature, these outcrops will constitute parallel ridges, or parallel lines of hills; when the region has been base-leveled, the outcrops will be in parallel belts, though no longer ridges or hills. The lower the plane of truncation, the farther apart the outcrops will be in the anticline, and the nearer together in the syncline (compare outcrop of *H*, along *ab* and *cd*, Fig. 117).

If, on the other hand, the axis of the anticline or syncline is not horizontal, that is, if it *plunges*, the topographic result will be different. In this case the outcrops of a given layer on opposite sides of an anticline will converge in the direction of plunge, and come together at the end. At a stage of erosion antedating peneplanation (say late maturity) there will be a ridge or succession of hills, in the position corresponding to the outcrop of a hard layer, with the canoe-shaped valley between. If two hard layers are involved, instead of one, there will be two encircling ridges, with a curved valley between them, and a canoe-shaped valley within the innermost (Fig. 127). A succession of plunging

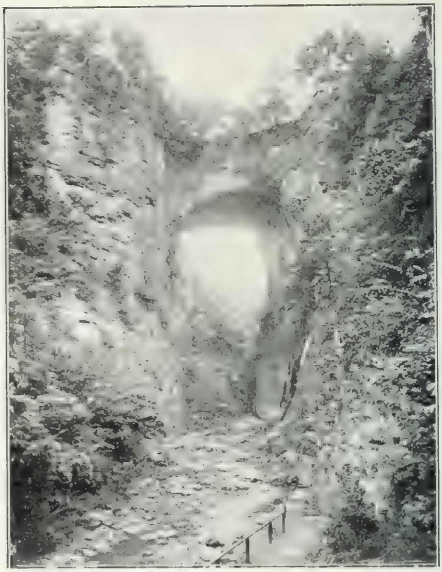


Fig. 125. — The Natural Bridge of Virginia. (U. S. Geol. Surv.)

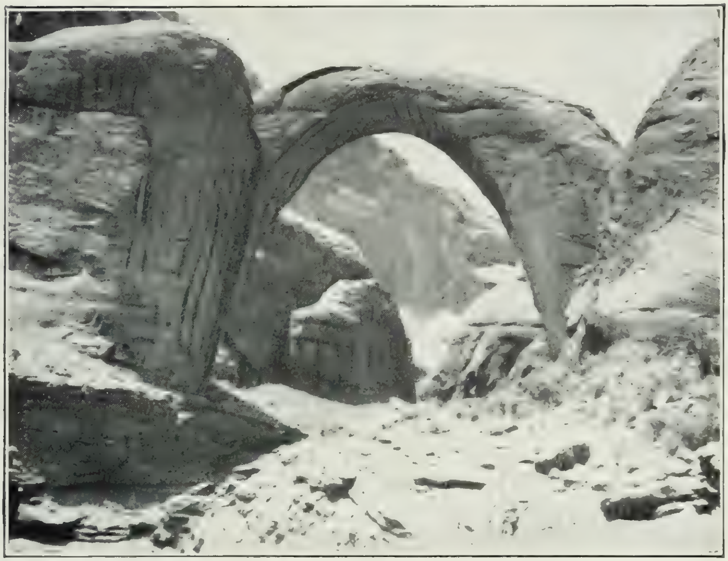


Fig. 126. — The Rainbow Bridge, San Juan Co., Utah. (Gregory, U. S. Geol. Surv.)

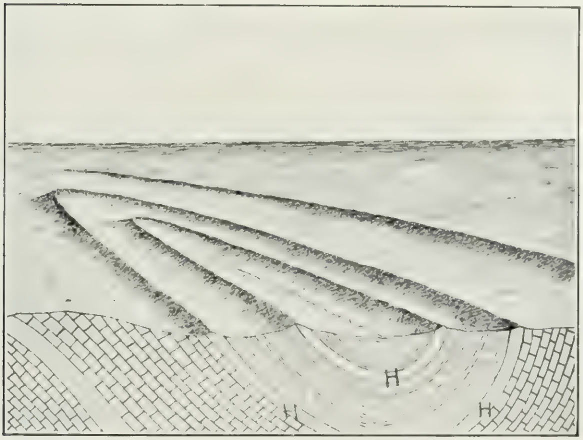


Fig. 127. — A canoe-shaped valley bordered by a ridge formed by the outcrop of a hard layer in a plunging syncline. The ridge bounding the canoe-valley is separated from an outer ridge by a curved valley, underlain by relatively weak rock. (Willis.)

anticlines and synclines might give rise to a very complex series of ridges and valleys. Illustrations of the above phenomena are found in various parts of the Appalachian Mountains.

In the structural adjustment which accompanies the erosion of folds, it often happens that the valleys come to be located in the eroded anticlines, while the outcrops of the hard layers on the flanks of such anticlines, or even in the original synclines, become the mountains (Fig. 128).

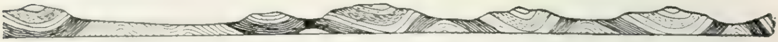


Fig. 128. — Cross-section of a portion of the Appalachian Mountains to illustrate the phenomena of erosion cycles. (Rogers.)

EFFECTS OF CHANGES OF LEVEL

Rise. If, after being base-leveled, or notably reduced by erosion, a region is uplifted so as to increase the gradients and velocities of its streams, it is said to be *rejuvenated*, and a new cycle of erosion is begun. The renewed youth differs from the first youth, in that the streams are already in existence in well-defined channels. The rejuvenated streams erode their valleys, after the manner of youthful streams. They excavate new valleys in the bottoms of older ones (Figs. 129 and 130), deepening, or tending to deepen them, to the new level of base-level.

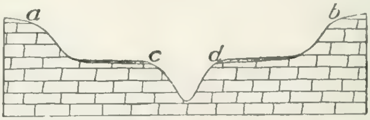
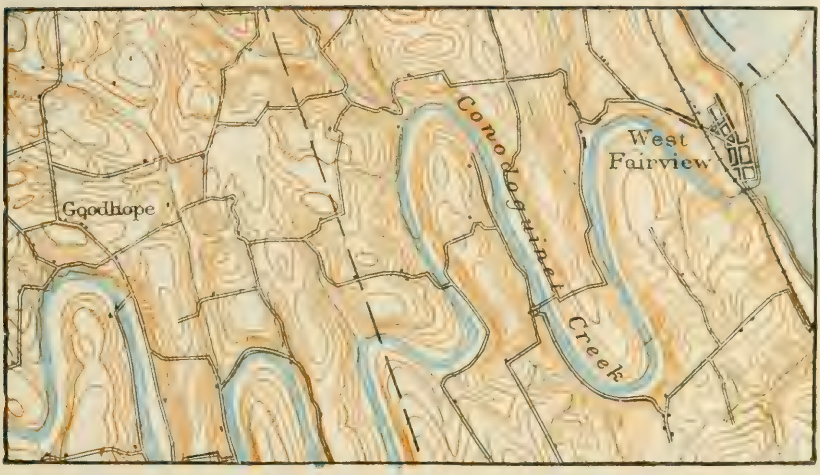


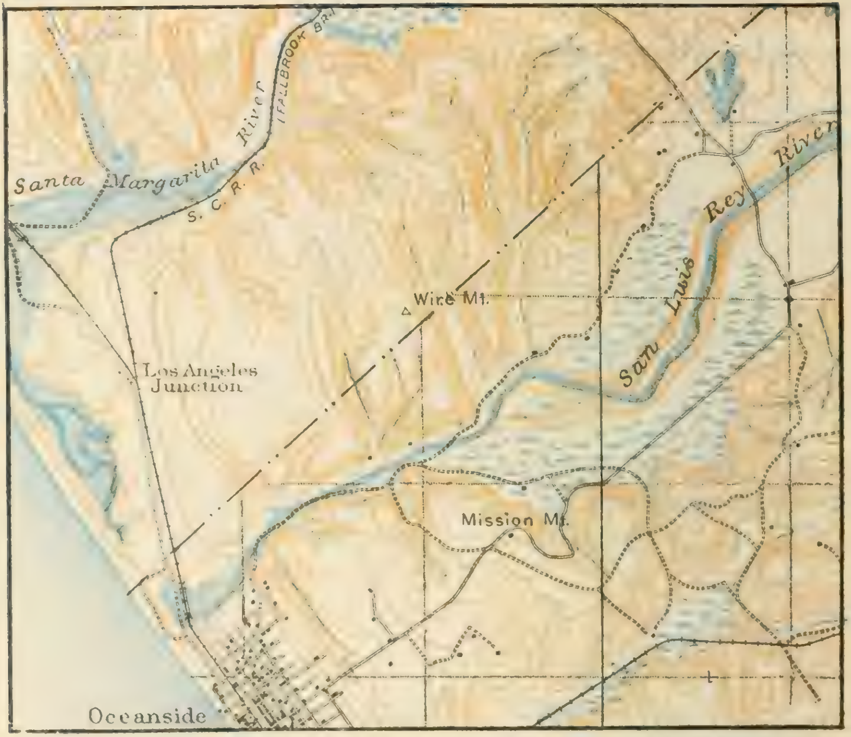
Fig. 129. — Cross-section of a wide valley, *ab*, in the bottom of which a younger valley, *cd*, has been excavated, as the result of uplift.

The new valley within the old one may develop along its whole course at the same time, or may begin at the


Goodhope
Conodoguinet Creek
West Fairview

Entrenched meanders. Scale, about 1 mile per inch. (Harrisburg, Pa., Sheet. U. S. Geol. Surv.)

PLATE XII


Santa Margarita River
S. C. R. R.
(FALLBROOK BR.)
Wire Mt.
Los Angeles Junction
San Luis Rey River
Mission Mt.
Oceanside

A Section of the California Coast, showing lands near the coast, which have recently emerged. Scale, about 1 mile per inch. (Oceanside, Cal., Sheet, U. S. Geol. Surv.)

debouchure of a stream and work headward. In either case, the tributaries are rejuvenated when their main is lowered at the point of union.

If an old winding stream is rejuvenated, the deepened channel follows the course which the stream had before rejuvenation. The result is that a new winding gorge is cut; that is, the old meanders are *entrenched.* Entrenched meanders, common in the Appalachian Mountains (Pl. XI), are known in many parts of the world (Fig. 131). Entrenched meanders and new valleys in the bottoms of old ones are among the commoner marks of rejuvenation. With rejuvenation of drainage, a new cycle of erosion is begun, whether the preceding one was complete or not.

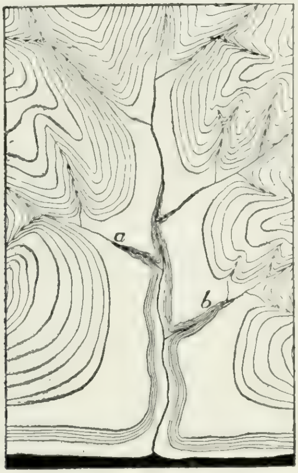


Fig. 130. — Diagram to illustrate an ideal case of rejuvenation as the result of uplift. The black area at the bottom represents the sea.

The recognition of successive cycles of erosion, separated by uplifts, is often easy. The principles involved are illustrated by Figure 132, which represents an ideal cross-section of considerable length (say 20 miles). The points *a*, *a′*, and *a″* have about the same elevation. Below them are the areas *b*, *b′*, *b″*, which have a nearly common eleva-

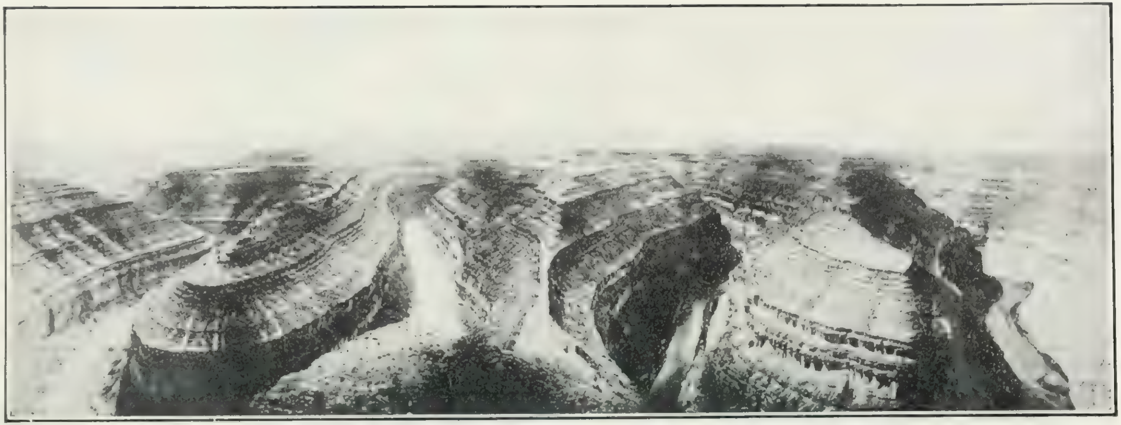


Fig. 131. — Entrenched Meanders. (Gregory, U. S. Geol. Surv.)

tion, below which are the sharp valleys *d*, *d′*, *d″*. The points *a*, *a′*, *a″* represent the cross-sections of ridges formed by the outcrops of layers of hard rock. If the ridges are level-crested the points of *a*, *a′*, and *a″* must represent remnants of an old base-level, *since at no time after a ridge of hard rock becomes deeply notched does it acquire an even crest, until it is base-leveled.* After the cycle represented by the remnants *a*, *a′*, and *a″* was completed, the region suffered uplift. A new cycle represented by the plain *b*, *b′*, and *b″* was well advanced, though not completed, when the region was again elevated, and the rejuvenated streams began to cut their valleys *d*, *d′*, and *d″* below the lowland plain of the previous incomplete cycle. The elevations, *c* and *c′* (intermediate in elevation between *a*, *a′*, and *a″*, and *b*, *b′*, and *b″*), may represent either remnants of the first base-leveled plain, lowered, but not obliterated, while the plain *b*, *b′*, and *b″* was developing; or they may represent a cycle intermediate between that during which *a*, *a′*, and *a″* and *b*, *b′*, and *b″* were developed.

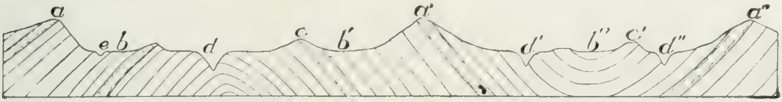


Fig. 132. — Diagram to illustrate cycles of erosion where the beds are tilted.

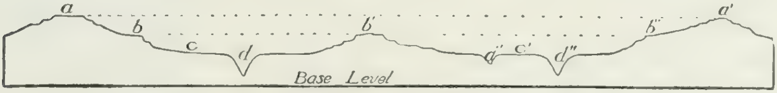


Fig. 133. — Diagram to illustrate cycles of erosion where the beds are horizontal.

If the strata involved are horizontal, the determination of the cycles is apt to be less easy. Thus in Figure 133, it is not possible to say whether *a* and *a′* represent remnants of an old base-level, or the original surface from which degradation started. So, too, the various benches below *a*, such as *b*, *b′*, and *b″*, may possibly be the result of superior hardness of beds at this level. For the determination in the field of successive cycles, it is necessary to consider areas of considerable size, and to eliminate from consideration the topographic effects of inequalities of hardness.

The application of the preceding principles makes it clear that the Appalachian Mountains, after being folded, were reduced to a peneplain (the Kittatinny peneplain) from the Hudson River to Alabama. The old peneplain surface is indicated by the level crests of the Appalachian ridges. The system was then warped (not folded) up, and in the new cycle of erosion which followed, broad plains were developed at a new and lower level, corresponding in a general way to the plains *b*, *b′*, and *b″* of Figure 132. The plains were located, for the most part, where the less resistant strata came to the surface. Above them rose even-crested ridges, the outcrops of the resistant layers, isolated by the degradation of the intervening weaker beds. It is the outcrops of these resistant layers which constitute many of the present mountain ridges (the high points of Fig. 132). The evenness of their crests testifies to the completeness of the first peneplanation. The evenness of the crests, is, however, interrupted by (1) notches cut by the streams in later cycles and (2) by elevations (monadnocks) above the common level. The monadnocks are mostly rather inconspicuous, but there is a notable group of them in North Carolina and Tennessee, of which Mount Mitchell and Roan Mountain are examples. When long distances are considered, the ridge crests depart somewhat from horizontality. This is believed to be due, in part at least, to the deformation of the old peneplain during the uplift which inaugurated the second cycle of erosion.

The extent to which the second cycle of erosion recorded in the present topography had proceeded before its interruption by upwarp, is indicated by the extent of the valley plains (Fig. 132) below the mountain ridges. While these plains were being developed on the weak rocks, narrow valleys only were cut in the resistant rocks which stood out as ridges.

The second cycle of erosion then, while still far from complete, was interrupted by uplift (relative or absolute), and a new cycle inaugurated. This event was so recent that the new (third) cycle has not yet advanced far.

Some of the features just described are illustrated by Figure 105. The even mountain crest in the background is the Kittatinny Mountain of New Jersey and its continuation in Pennsylvania. In common with other corresponding crests, it is a remnant of the oldest recorded base-level (or peneplain) of the region. The great gap in the mountain is the Delaware Water-Gap. Below the mountain crest there is another plain, developed in a subsequent cycle of erosion, while the valley plain in the foreground represents the work of a still later cycle.

Many peculiarities of the drainage of the Appalachian Mountain system are intimately connected with the history just outlined. Thus three great rivers, the Delaware, the Susquehanna, and the Potomac, have their sources west of the Appalachians proper, cross the system in apparent disregard of the structure, and flow into the Atlantic. The James and Roanoke head far to the west, although not beyond the mountain system, and flow eastward, while the New River (leading to the Kanawha), farther south, heads east of the mountain-folds, and flows northwest across the alternating hard and soft beds of the whole Appalachian system, to the Ohio (Fig. 134). The French Broad, a tributary of the Tennessee, has a similar course. The courses of these streams were apparently assumed during the time of the Kittatinny base-level, when the region was so flat that streams could flow irrespective of structure. Elevation rejuvenated them, and they have held their courses in succeeding cycles across beds of unequal resistance, though smaller streams have become more nearly adjusted to the structure. Crustal deformations have also helped these rivers to hold their courses, for the Kittatinny peneplain seems to have been tilted to the southeast at its northern end, and to the southwest at its southern when the succeeding cycle began.

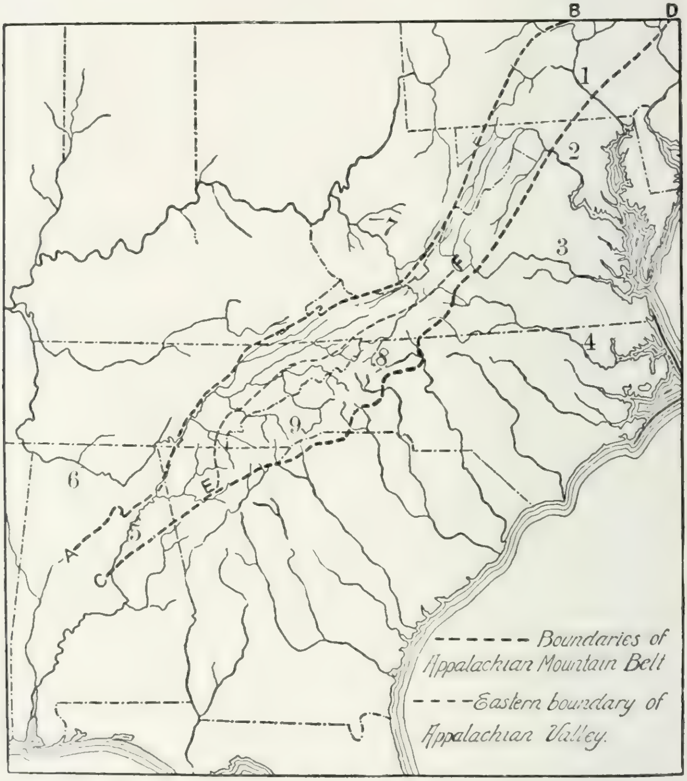

B
D
1
2
3
4
5
6
7
8
9
A
C
E
F
Boundaries of Appalachian Mountain Belt
Eastern boundary of Appalachian Valley.

Fig. 134. — Showing certain peculiarities of Appalachian drainage. 1 = the Susquehanna; 2 = the Potomac; 3 = the James; 4 = the Roanoke; 5 = the Coosa; 6 = the Tennessee; 7 = the Kanawha; 8 = head of New River; 9 = head of the French Broad.

Streams which hold their early courses in spite of changes which have taken place since their courses were assumed, are said to be *antecedent*. They antedate the crustal movements which, but for pre-existent streams, would have given origin to a different arrangement of river courses. As a result of crustal movements, therefore, a consequent stream may become antecedent. Master streams are more likely than subordinate ones to hold their courses, and therefore to become antecedent.

The uplift of base-leveled beds, especially if the beds are tilted so as to bring layers of unequal resistance to the surface at frequent intervals, affords conditions favorable for extensive adjustment. The numerous wind-gaps in the mountain ridges, representing the abandoned courses of minor streams, and the less numerous water-gaps, which indicate the resistance of large streams to structural adjustment, are instructive witnesses of the extent to which adjustment has taken place. So extensive has been the adjustment among the streams of the Appalachian Mountains that there is probably no considerable stream in the whole system which has not gained through its own piracy or lost through that of its neighbors.

Sinking. The land on which a river system is developed may be depressed relative to sea-level. In this case the sea occupies the lower ends of valleys, converting them into bays and estuaries. A valley in this condition is said to be *drowned*. Of drowned valleys there are many examples along the Atlantic coast; thus, for example, the St. Lawrence is drowned up to Montreal, and the Hudson up to Albany. If the drowned portion of the Hudson valley were not so narrow, it would be a bay. Delaware and Chesapeake bays, as well as many smaller ones, both north and south, are likewise the drowned ends of river valleys (Figs. 135 and 136).

Successive Rising and Sinking. Another peculiarity of valleys and streams resulting from changes of level is illustrated by Plate XII. The main valleys of this part of the coast were developed when the land stood considerably higher than now. Later, the subsidence of the coast converted the lower ends of the valleys into bays or *fiords*. The bays were then transformed into lakes by deposition at their mouths. Subsequent rise of the land or depression of the sea allowed the drainage from the old lagoons to cut across the deposits which had converted the bays into lagoons. The result in each case is an old, wide valley above, succeeded by a young one near the debouchure. The low sandbars across the present mouths are recent features.

Differential Movement. Warping. A land surface on which a river system is established may suffer warping, some parts going up and others down. Above an upwarp which notably checks its flow, a stream is *ponded*. If a stream holds its course across a notable uplift athwart its valley, it becomes an *antecedent stream*. The Columbia River is thought to have held its antecedent course across areas which have been uplifted (differentially) hundreds and even thousands of feet. Some striking scenic features of this

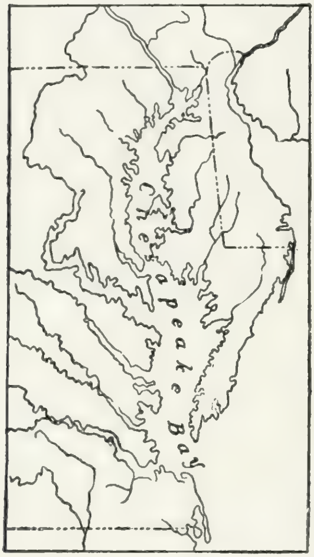


Fig. 135

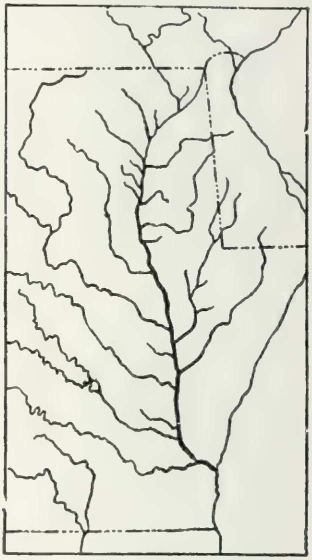


Fig. 136

Fig. 135. — Chesapeake Bay and its surroundings. The bay is a drowned river valley, and the lower ends of its tributary valleys are also drowned.

Fig. 136. — The drainage of the region about Chesapeake Bay as it would have been but for drowning.

noble valley are the result of these changes in the country through which it flows. A lesser stream would have been diverted, as many of the Columbia's tributaries have been. Even its course across the Cascade Range has been interpreted as antecedent.

THE AGGRADATIONAL WORK OF RUNNING WATER

We have seen that rivers carry mud, sand, gravel, etc., from land to sea, and that their goal is the degradation of the land nearly to the level of the sea. We have also seen that rivers do not always carry the sediment derived from the land directly to the sea; that instead it is often dropped for a time on the land, perhaps to be picked up and carried on again when the conditions for its transportation are more favorable. We have now to inquire more particularly into the causes and especially the results of deposition.

CAUSES OF DEPOSITION

When running water loses its load, or any part of it, it is generally because the current has lost some of its velocity and is no longer able to sweep up as much sediment as settles out. We have already seen that gradient and volume are the most important factors in determining the velocity of a small stream.

The commonest cause of loss of velocity is decrease of gradient. Running water may lose velocity (1) suddenly, as when it passes from a steep slope, whether of hill or mountain, to a gentle one, or to a body of standing water, or a more slowly moving stream, or (2) slowly, as in descending a valley whose gradient becomes gradually less. We therefore look to the places where these changes in velocity occur for the principal deposits of running water. Streams also become slower wherever their channels become wider, even if volumes and gradients remain constant.

Decrease of volume is a less common cause of decrease of velocity. Streams generally increase in size with increasing distance from their sources, but to this general rule there are exceptions: (1) If a stream flows through a very dry region, it may receive few tributaries and few springs, while evaporation is great and the thirsty soil and rock through which it flows absorb some of the water. In a dry region, therefore, a stream may diminish as it flows, and may even disappear altogether (Pls. XIII and II). (2) A stream sometimes breaks up into several streams (Fig. 139), the volume of each being less than that of the original stream. (3) Again, many streams, especially in arid regions, have much of their water withdrawn for purposes of irrigation. Many streams in the West lose volume in this way. (4) Streams decrease in volume as their floods decline, and so deposit in their channels during low water, where they eroded during high water.

LOCATION OF ALLUVIAL DEPOSITS AND THEIR TOPOGRAPHIC FORMS

1. At the Bases of Steep Slopes. Every shower washes fine sediment down the slopes of the hills, and much of it is left at their bases. Its lodgment is shown by the fact that fences in such situations are sometimes buried by the mud lodged against them. Temporary streams, bred of showers, sometimes flow down steep slopes, and are suddenly checked at the bases, gathering much debris from the steep slopes but abandoning it where their velocity is suddenly checked. Thus, at the lower end of a new-made gully on a hillside there is commonly a mass of debris which was washed out of the gully itself (Fig 63). Material in such positions accumulates in the form of a partial cone, known as an *alluvial cone*, or if of gentle slope an *alluvial fan.*

Alluvial cones have much in common with cones of talus; but in the latter, gravity brings the material down without the aid of water, or with but little help from it. Between talus cones and alluvial cones there is, however, a complete gradation. Conspicuous alluvial cones are common at the bases of steep slopes in semi-arid regions; for in such regions the rainfall is fitful, and the occasional heavy showers, which give rise to temporary powerful torrents, favor the development of cones of great size. At the bases of the mountain ranges in the Great Basin of Utah and Nevada, the talus and alluvial cones from the mountains are in places 2000 or 3000 feet high.

An *alluvial fan* is similar to an alluvial cone, except that it has a lower angle of slope (Fig. 137). The term *fan* is indeed more appropriate than *cone* for most alluvial accumulations at the bases of slopes. The lower angle of the fan may be due to the less abrupt change of slope where it is developed, to the larger quantity of water concerned in its deposition, to the smaller amount of detritus, or to its greater fineness. Less change of slope, more water, and less and finer material, all favor the wider distribution of the sediment, and so the development of fans rather than cones. Nearly all young rivers descending from mountains build fans where they leave the mountains. For example, the rivers descending from the Sierra Nevada Mountains to the great valley of California build extensive fans at the base of

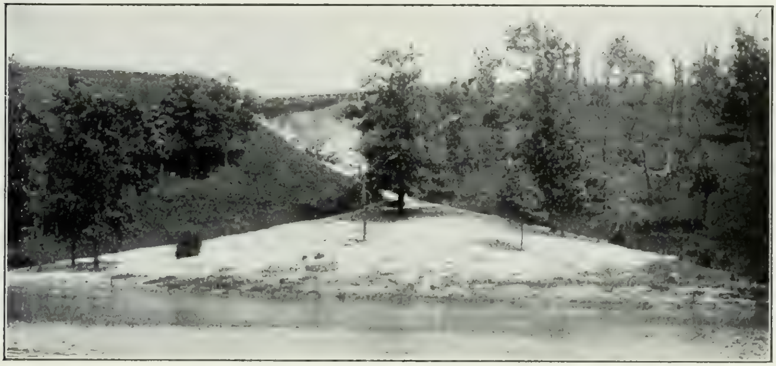


Fig. 137. — An alluvial fan. (U. S. Geol. Surv.)

the range. The fan of the Merced River has a radius of about 40 miles. Most of the rivers descending from the Rockies to the plains on the east have also built large fans.

The fans made by neighboring streams may grow laterally until they merge. The union of several such fans makes a *compound alluvial fan*, or a piedmont alluvial plain (Pl. XIII). Such plains exist at the bases of most considerable mountain ranges. Sheet wash, as well as streams, contribute to them. The depth of alluvial material in such plains may be scores and even many hundreds of feet. One of the most remarkable features of these land deposits is their great spread. Thus east of the Rocky Mountains they extend out more than a hundred miles in some places. This wide spread appears to be the result of the long-continued action of running water. The alluvial cone or steep fan, as first built, is degraded later, and its materials spread more widely. As the gradient of the running water which makes the fan becomes lower, the higher parts of the fan are spread out farther and farther, with continually lower gradient.

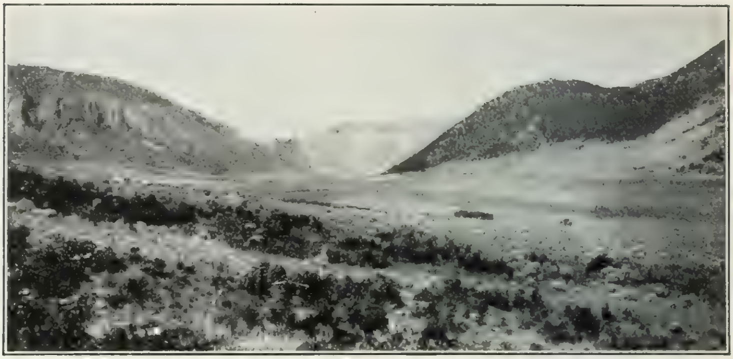
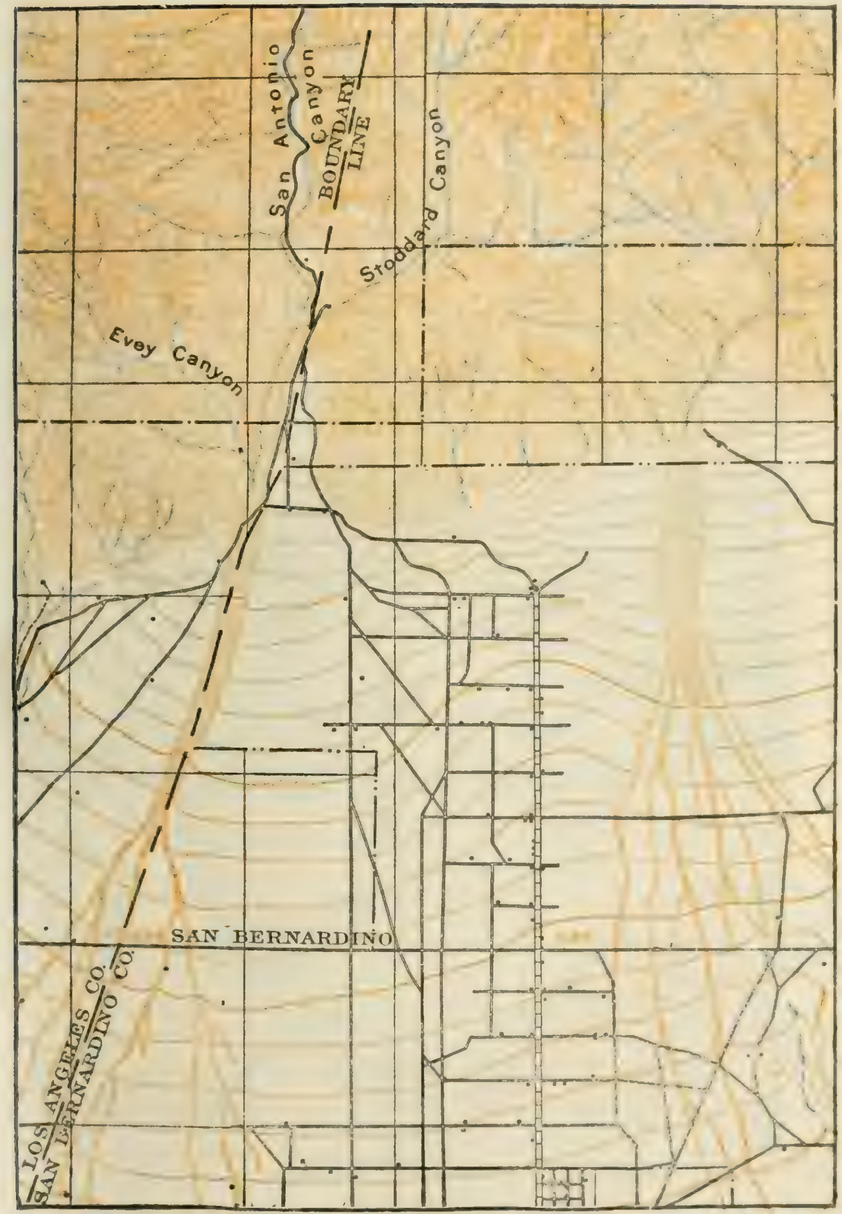


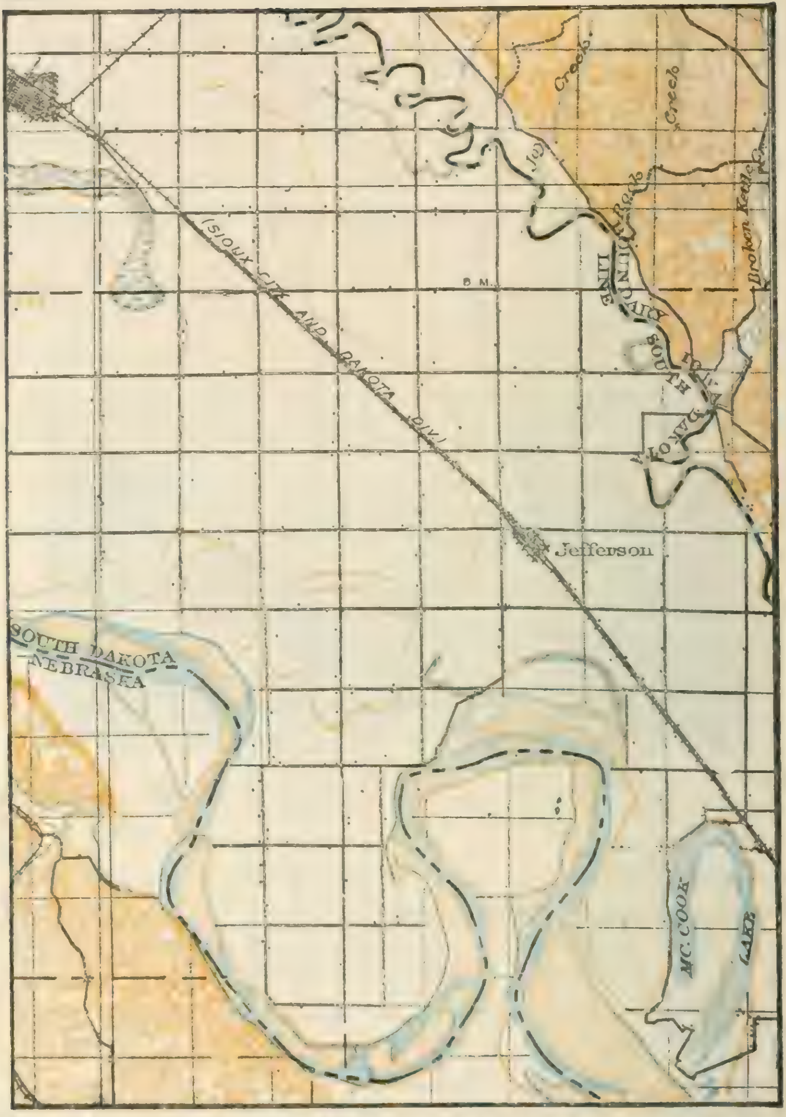
Fig. 138. — Valley showing deposition at the bases of valley slopes, tending to give the valley a U-shaped base. Unaweep Canyon, Colo. (Cross, U. S. Geol. Surv.)

Probably deposits of this sort have been far more important in the past than has been generally recognized. Much of the material of the Coastal Plain of the Atlantic and the


San Antonio Canyon
BOUNDARY LINE
Stoddard Canyon
Evey Canyon
SAN BERNARDINO
LOS ANGELES CO.
SAN BERNARDINO CO.

A piedmont alluvial plain or compound alluvial fan in Southern California. Scale about 1 mile per inch. (Cucamonga, Cal., Sheet, U. S. Geol. Surv.)

PLATE XIV


Jay
Creek
Creek
Rock
Broken Kettle Cr.
IOWA
SOUTH DAKOTA
LINE
B.M.
(SIOUX CITY AND DAKOTA DIV.)
Jefferson
SOUTH DAKOTA
NEBRASKA
MC COOK
LAKE

Stream flats. The Missouri and Big Sioux Rivers. Scale, about 2 miles per inch. (Elk Point, S. Dak.—Ia.—Neb. Sheet, U. S. Geol. Surv.)

Gulf slopes of the United States appears to have been deposited in this way. A large part of the Great Plains is covered with wash from the Rocky Mountains, and similar deposits are of great extent and depth east of the Andes and south of the Himalayas. They are, indeed, of significant extent and depth about almost every mountain range which has been studied, especially where plains are adjacent to the mountains. It seems clear that similar deposits must have been made at all stages in the past history of the earth, whenever and wherever mountainous lands bordered plains. Formations of this general sort, made at the bases of high lands, have now been recognized among the ancient, as well as the recent, formations of the earth, and some of the ancient beds of sediment deposited in this way are hundreds and even thousands of feet thick. They probably attained their greatest thickness, as now, in basins. Since deposited on the land, they are *terrestrial formations* of *fluvial* origin.

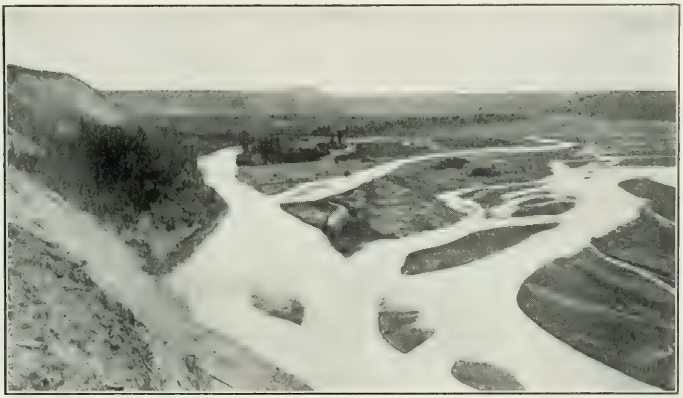


Fig. 139. — A branching stream. Junction of the Cooper and Yukon rivers, Alaska. Shows also bars, etc. (U. S. Geol. Surv.)

2. **In Valley Bottoms.** A stream which makes deposits in its channel reduces the size of its channel, which, in time, may become too small to hold all the water. A part then breaks out, and follows a new course over the valley flat.

This process may be repeated again and again, and the stream becomes branching or *braided* (Fig. 139). Streams sometimes deposit bars in their channels, especially at low water. The bars may be swept away in time of flood, but they sometimes become more or less permanent islands.

The profiles of most valleys are curves, the curvature becoming less as the lower end of the stream is approached (Fig. 140). It therefore happens that as a stream descends its valley it generally reaches a point where its reduced gradient so diminishes its velocity that, in spite of its increased volume, it must abandon some of its load. In this way sediment is distributed for long distances along valley bottoms. It is left in the channels of streams in low water, and spread over their flood-plains in time of flood, aggrading them and making them alluvial plains. Deposition in a valley which has no flat tends to develop one (Fig. 141). Alluvial deposits on valley flats are usually but a few feet, or a few scores of feet thick; but in some cases they reach hundreds of feet. It is evident that flood-plains in normal river valleys are commonly produced chiefly by erosion and that the veneer of alluvium in most cases is of little consequence to the shape of the valley.

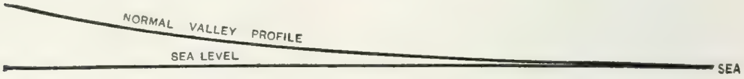


Fig. 140. — Profile of a normal valley.

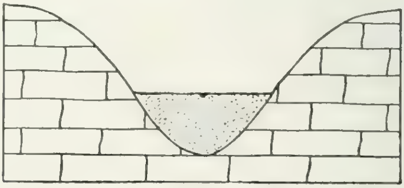


Fig. 141. — Flat developed by aggradation — diagrammatic.

At times of occasional floods, the current in the main channel is swift; but as the water escapes its channel and spreads over the adjacent flat, its velocity is promptly checked, because its depth is reduced while friction upon the bed, vegetation, etc., becomes greater. It, therefore, then and there abandons much of its load. Repeated deposition in this position, in excess of that over other parts of the flood-plain, gives rise to low ridges along the stream banks known as *natural levees.*

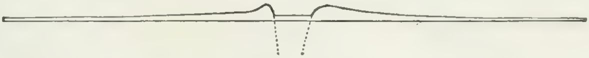


Fig. 142. — Levees of the Mississippi in cross-section, four miles north of Donaldsonville, La. Vertical scale x 50. The horizontal line represents sea-level.

Scour-and-fill. In time of flood when the velocity is high, aggrading streams dig out their channels deeply, only to fill them up again as the flood subsides and velocity diminishes. This alternate deepening and filling is known as *scour-and-fill.* It is well illustrated by the Missouri River. At Nebraska City, scour reaches depths of 70 to 90 feet occasionally. At Blair, about 25 miles above Omaha, the

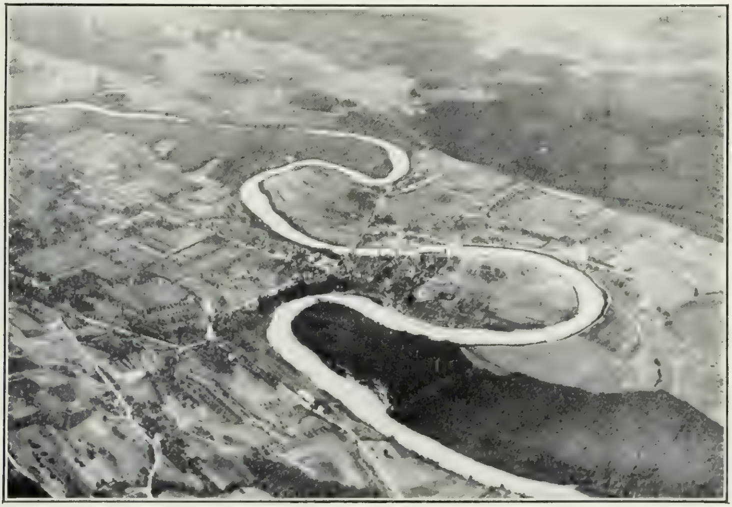


Fig. 143. — Meandering river, Brungarten, Switzerland. (Courtesy Swiss Air Serv.)

same river is believed to cut to bed rock (about 40 feet below the bottom of the channel in low water) during floods. All streams similarly situated do a like work. The material thus eroded is shifted downstream, some of it for short distances only, and some of it to the sea. Even an aggrading stream therefore is not without erosive activity. It is a stream whose fill exceeds its scour, not one which has ceased to erode. A stream when in flood simultaneously aggrades its plain and degrades its channel.

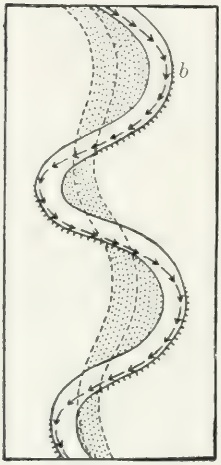


Fig. 144

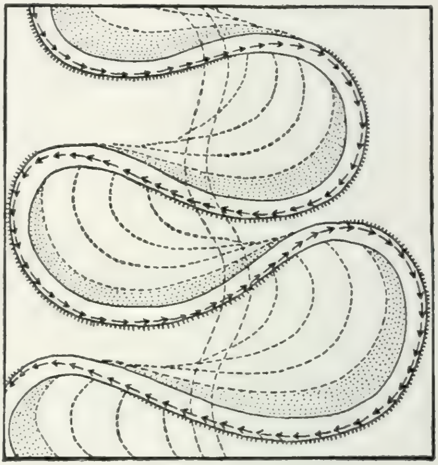


Fig. 145

Fig. 144. — Diagram illustrating an early stage in the development of river meanders. The dotted area represents the area over which the stream has worked.

Fig. 145. — A later stage in the development of meanders.

Materials of the Flood-plain. As a result of its varying velocities during flood and low water, a stream may deposit coarse material at one *time* and fine at another; also coarse at one *place* and fine at another. Flood-plain deposits are therefore characteristically very heterogeneous, ranging from the finest mud, through sand, to gravel and even boulders. In general they become finer downstream. In a given plain they are normally coarser at the bottom, and finer toward the top, though stratified throughout.

Flood-plain Meanders. Any stream on a flood-plain is likely to meander widely (Pls. XIV and VIII and Fig. 143). In general terms this may be said to be the result of low velocity, which allows the stream to be turned aside easily. Were the course of such a stream made straight, it would soon become crooked again. The manner of change is illustrated by Figures 144 and 145. If the banks are less resistant at some points than at others, as is always the case, the stream will cut more at such points. If the configuration of the channel is such as to direct a current against a given point (*b*, Fig. 144), the result is the same, even without inequality of material. Once a curve in a bank is started, it is increased by the current which is directed into it.

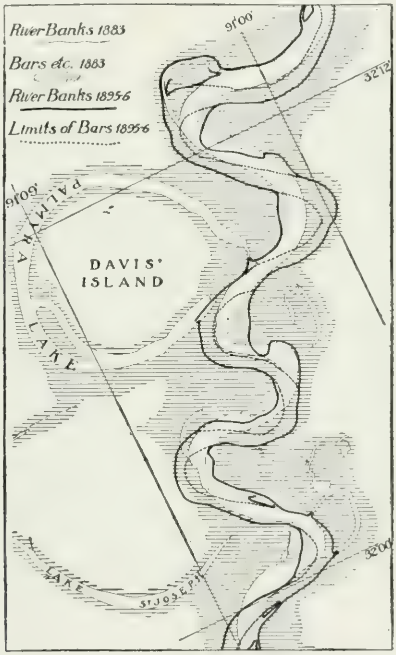

River Banks 1883
Bars etc. 1883
River Banks 1895-6
Limits of Bars 1895-6
9100'
3212
PALMYRA
DAVIS' ISLAND
LAKE
LAKE ST JOSEPH
3200

Fig. 146. — Meanders and cut-offs in the Mississippi Valley below Vicksburg. The figure shows the migration of the meanders downstream, and their tendency to increase.

Furthermore, as the current issues from the curve, it impinges against the opposite bank and develops a curve at that point. The water issuing from this curve develops another, and so on.

Once started, the curves or meanders tend to become more and more pronounced (Fig. 145). In the case represented by Figure 1, Plate VIII, the narrow neck of land between the curves is almost cut through. A later stage in the process is shown in Figure 2, Plate VIII.

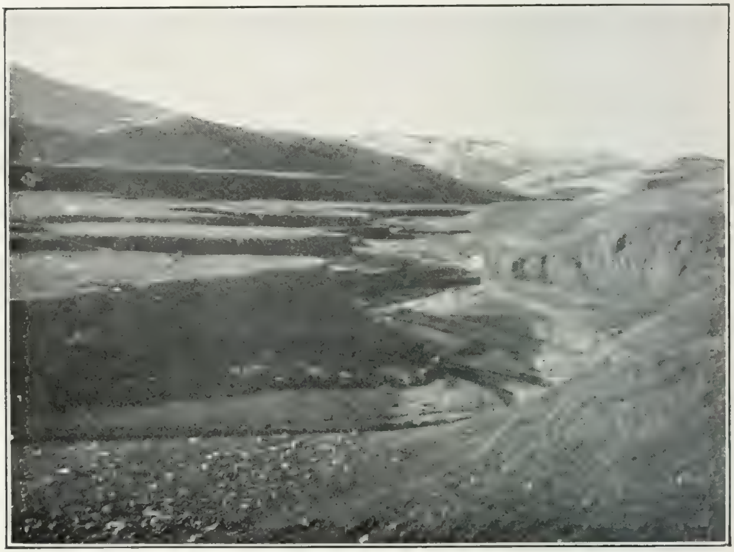


Fig. 147. — Alluvial terraces on Santiago creek, Calif. (Pack. U. S. Geol. Surv.)

When the stream has cut off a meander, the abandoned part of the channel may remain unfilled with sediment. If it contains standing water, as is true in many cases, it becomes the site of a lake or *bayou* (Fig. 146). If crescentic in shape, such lakes are called *ox-bow* lakes (Pls. XIV and VIII).

Alluvial Terraces. Stream terraces are bench-like flats or narrow plains along the sides of valleys and above their bottoms (Fig. 147). They are, as a rule, narrow but some have great length in the direction of the axis of the valley.

The commoner river terraces are remnants of former flood-plains below which the streams have cut their channels. They originate in various ways: (1) If an alluvial flood-plain has been built as the result of an excessive supply of sediment, the exhaustion or diminution of the excessive supply would leave the stream again relatively clear and free to erode where it had been depositing. It would forthwith set to work to carry away the material which it had temporarily deposited on the plain. The plains built up in many valleys in the northern part of our continent during the glacial period, when the drainage from the ice coursed through them, have subsequently been cut into by erosion, and their remnants have become terraces. (2) A notable increase in the volume of a stream, without corresponding increase in load, as when one stream captures another, may occasion the development of terraces by allowing the enlarged stream to deepen its channel. (3) The uplift of a region in which there are well-developed river flats, would

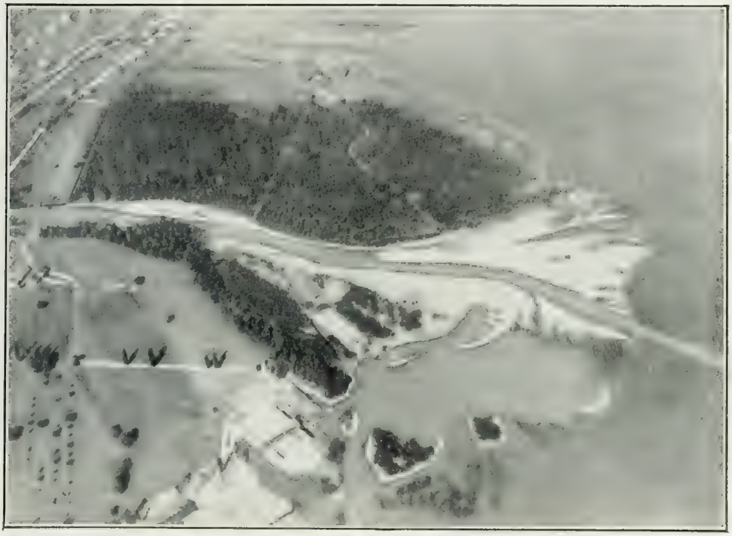


Fig. 148. — Delta of the Kander River, Switzerland, from an elevation of 1500 feet. (Swiss Air Serv.)

rejuvenate the streams, and parts of their old flood-plains would be left as terraces. Other causes, outside the scope of this discussion, may give rise to terraces.

3. At Debouchures. Where a stream flows into the sea or a lake, its current is promptly checked and soon destroyed altogether. Its load accordingly settles out and accumulates on the bottom. If not washed away by the waves or currents, deposits of river-borne sediment in such places make *deltas* (Figs. 148 and 149).

The delta has some points in common with the alluvial fan. In both cases the principal deposit is concentrated at the point where the velocity is suddenly checked. In the case of the delta, however, the current is checked more completely, and consequently the delta differs from the alluvial fan in that its outer edge has a steep frontal slope (compare Figs. 150 and 151).

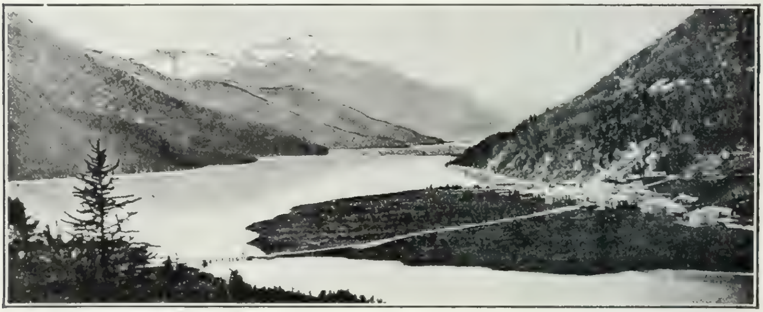


Fig. 149. — A delta in a lake. The village is Silva Plana, in the Engadine, Switzerland. (Robin.)

Once a delta is started below the water, deposition takes place upon its surface, which may be built up to, and even above, the water-level. That part of the delta above the surface of the water in which it is built, is like a flat alluvial fan. The growth of a delta determines the attitude of its stratification (i.e., the structure). At any stage in its growth the river discharges its sediment across that part of the platform already built. The sediment rolled at the bottom of the current is dumped on reaching the steep slope and constitutes the inclined *fore-set* beds shown in Figure 150. Material in suspension is carried farther, settles more gradually, and constitutes the *bottom-set* beds (*c*, Fig. 150). In time the bottom-set beds, originally deposited some distance beyond the debouchure, may come to be overlain by the fore-set beds deposited at a later time. While the fore-set beds are being deposited on the steep frontal slopes of the delta, and the bottom-set beds beyond, deposition is also taking place on the top of the delta, as in an alluvial fan. These *top-set* beds are laid down in a nearly horizontal position, and their seaward margin is gradually extended. Thus the delta comes to have the three-fold structure shown in Figure 150.

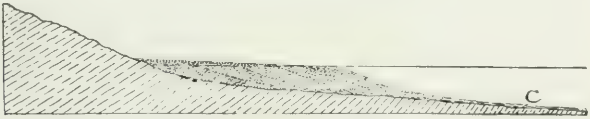


Fig. 150. — Diagrammatic profile and section of a delta.

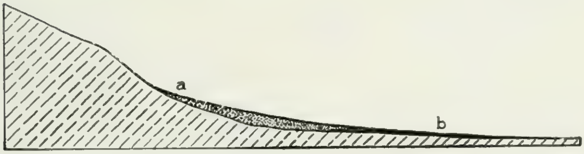


Fig. 151. — Diagrammatic profile and section of an alluvial fan.

Much land has been made by delta building. Thus the Colorado River has built a great delta many square miles in area at the head of the Gulf of California (Fig. 152). The delta has been built quite across the gulf near its upper end, shutting off the head. In the arid climate of this region, this shut-off head has become a nearly dry basin, the lowest part of which is about 300 feet below sea-level. The Skagit River, in Washington, has built out its delta so as to surround high islands in Puget Sound, thus joining them to the mainland. The deltas of the Mississippi (Fig. 153), the Nile and the Hoang-Ho rivers are among the large and well-known deltas of the world. The united delta of the Ganges and Brahmaputra is also a great one, having an area (above water) of 50,000 square miles. The Po has built a delta 14 miles beyond the former port of Adria, which gave its name to the Adriatic Sea, while the Rhone has advanced its delta some 15 miles in as many centuries.

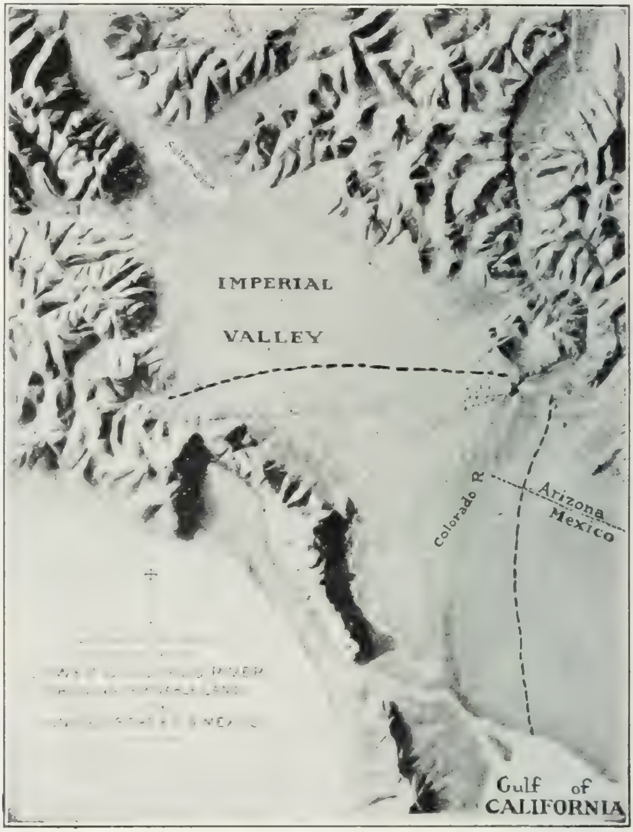


Fig. 152. — Relief-map of an area about the head of the Gulf of California, showing the delta of the Colorado River, outlined, in a general way, by dotted lines. The Salton Sink is shown at the north, and the Imperial Valley lies south of the sink. (U. S. Rec. Sur.)

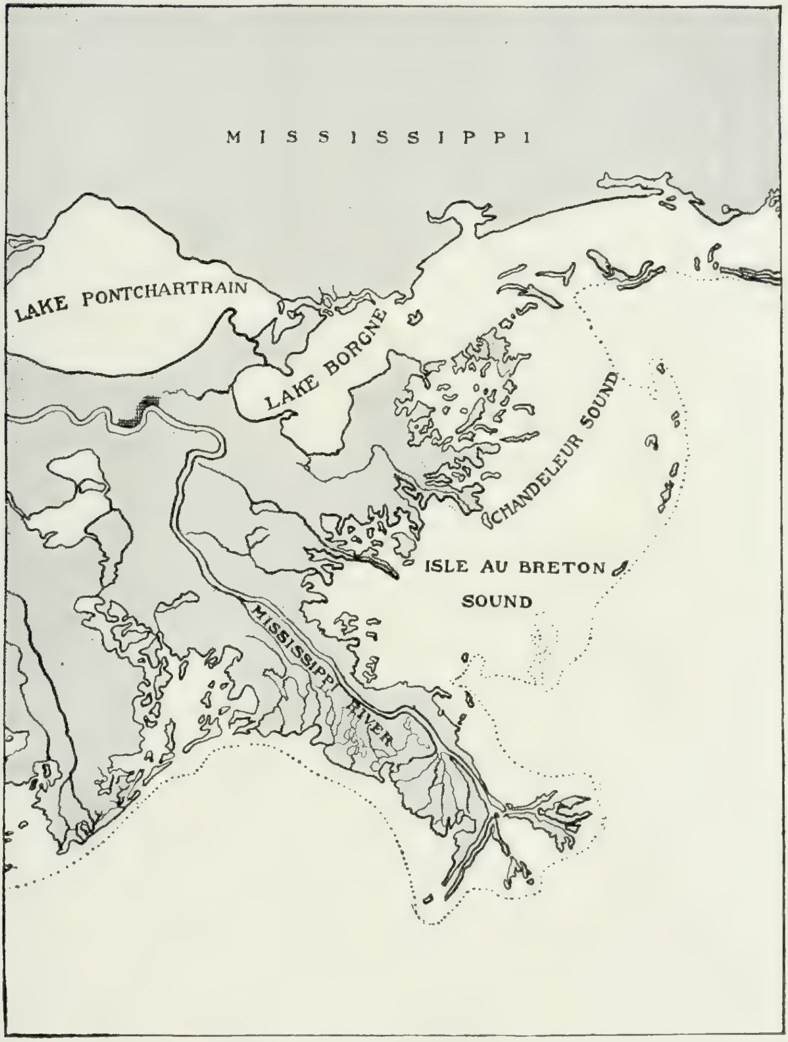

MISSISSIPPI
LAKE PONTCHARTRAIN
LAKE BORGNE
CHANDELEUR SOUND
ISLE AU BRETON
SOUND
MISSISSIPPI RIVER

Fig. 153. — Delta of the Mississippi. The dotted line outside the land represents the 3-fathom line.

The effect of delta-building is to increase the area of the land; but it is to be noted that the processes which lead to delta-building reduce the volume of the land-masses above sea-level, even though they increase their area.

The outlines of some deltas are determined by the configurations of coast-lines where they are built. When, for example, a delta is built into a bay, the form of the bay-head determines the shape of the delta. The normal form of a delta built on an open coast is somewhat semicircular, though there is often a fringe of *delta fingers* the whole having some resemblance to the Greek letter Δ, which fact suggested the name given to these terminal deposits of streams.

CHAPTER V

SNOW AND ICE

We have seen that the atmosphere, the ground-water, and the waters on the surface of the land bring about important changes in its configuration. Another agent of change is ice in its various forms.

Ice Beneath the Surface. The wedge-work of ice in the crevices of rock has already been mentioned (p. 42). When the great areas where water freezes during some part of the year are considered, it is clear that the aggregate effect of the freezing of water in the pores and crevices of rock, over long periods of time, must be very great. The water which freezes in the soil loosens and moves the surface portion. This is shown by the disturbance of the walls of buildings if their foundations do not go below the depth of freezing, by the lifting of fence posts, and by the working up of stones and boulders through the soil of the fields, as freezing and thawing succeed each other. The frozen water in the soil makes it solid, and temporarily retards or prevents surface erosion.

Ice on Lakes. To understand how ice is formed on lakes and ponds, it is necessary to follow the physical changes which take place in water and ice during the winter.

It is an important fact that water is densest, and hence heaviest, at a temperature of about 39° F. It obviously follows that it is lighter at temperatures both above and below this figure. In middle latitudes during the summer not only the surface water, but that at all depths to the bottom, in all but the deepest lakes, is above 39° F. As the surface water cools in autumn and winter, it sinks, warmer water taking its place. This warmer water is cooled in turn and sinks. Normally, this process continues until all the water of the lake is of the same density (maximum) and hence is at 39° F. With further cooling, the surface water expands somewhat, and so remains at the surface. When the surface water has cooled to 32° F., it freezes, and in freezing expands about one-tenth its volume.

Deep lakes in middle latitudes, such as the Great Lakes, do not freeze over even in the coldest weather, because the mass of water is not reduced to 39° F., and so long as the body of water is above its maximum density the cool surface waters continue to sink and the lake does not freeze over. Where the water is shallow around the margins of such lakes, ice forms in considerable abundance.

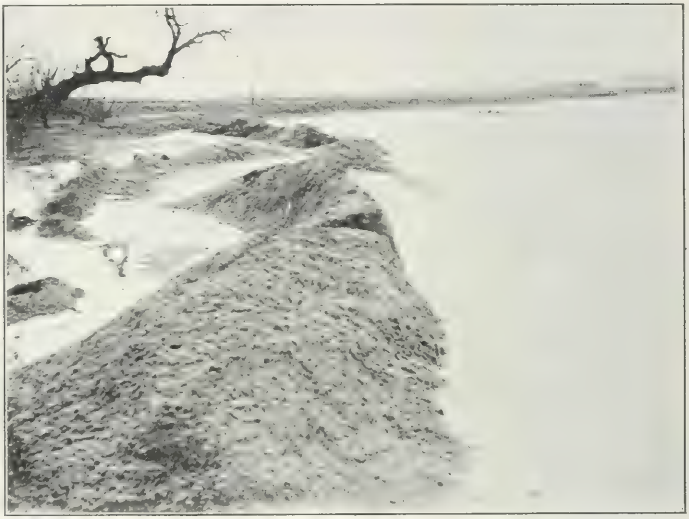


Fig. 154. — Shove of shore ice where the shore is marshy. The ice of the frozen marsh is pushed up into ridges. (Buckley, Wis. Geol. Survey.)

Ice, like other solids, contracts as its temperature falls. So if a lake is frozen over and the temperature later falls notably, the ice contracting pulls away from the shore, or cracks open. Water, exposed by these movements around the margins and in cracks, freezes promptly and the lake is again completely covered with a solid sheet of ice. When the cold "spell" is over and the temperature rises, say from —10° F. to 20° F., the ice expands, and the sheet of ice, becoming too large for the lake, must either crowd upon the shore or arch up in places.

If the water near the shore is very shallow, the ice freezes to the sand, gravel, and boulders at the bottom. If the shore of the lake is low, the ice may shove this material up year after year, eventually making a low ridge or "wall" slightly above the water level. Such "walled lakes" are not uncommon in the northern part of the United States.

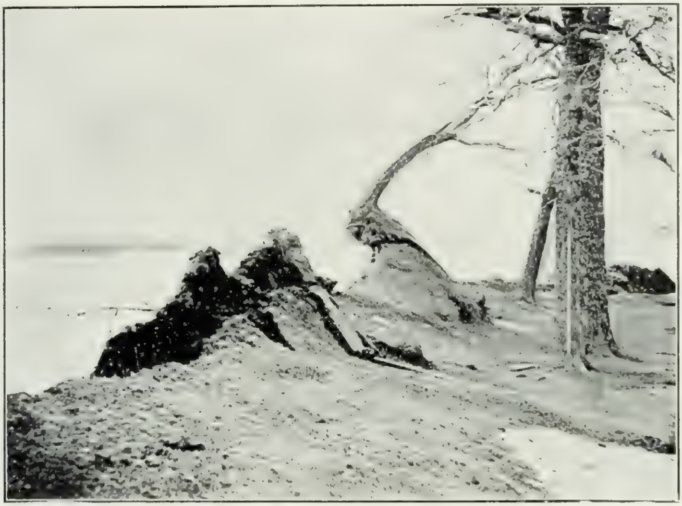


Fig. 155. — The shove of ice on the shore of Lake Mendota, Wis. (Buckley.)

If the lake is bordered by a low marsh, the ice and frozen earth of the marsh are really continuous with the ice of the lake, and the ice-push in places produces ridges in the marsh (Fig. 154). A succession of colder and less cold periods may give rise to a succession of such ridges. If the shore is steep, the crowding of the ice against a low cliff of yielding material, such as clay, may even overturn trees (Fig. 155).

Ice on Rivers. Rivers also freeze over in cold climates, and when the ice breaks up in the spring, the stones and boulders to which it was frozen are sometimes rafted miles down the river. At Montreal stone buildings 30 to 50 feet square, projecting so as to have river-ice form about them, have been moved by the ice of the St. Lawrence.

When river-ice breaks up, masses of it are carried downstream and may accumulate in vast fields or "jams" against obstructions in the river. Where a jam has formed above a bridge, the bridge may be swept away. The jams also occasion disastrous floods above their sites, and when they break, the waters accumulated above may sweep down the valleys with destructive violence.

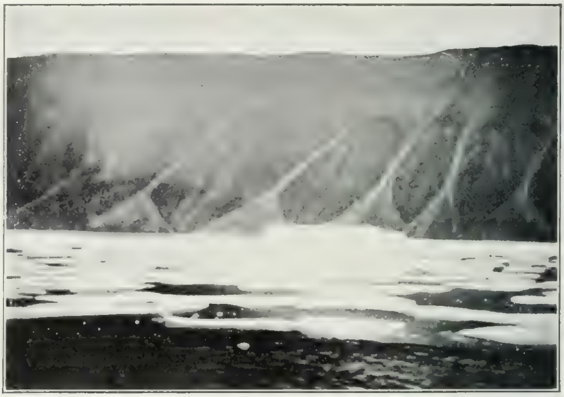


Fig. 156. — Floe-ice on the shore of Greenland.

Northward-flowing rivers in the northern continents are especially subject to such floods. The snows of their upper basins often melt while the lower parts of the streams are still frozen over. The free discharge of the upper waters is thus prevented for a time, and freshets follow later.

Ice on the Sea. In high latitudes, ice is formed along the seashore. Unlike fresh water, sea-water condenses until it freezes, at a temperature of 26° to 28° F., the variation being due to the varying degree of salinity of the water.

In polar regions the sea-ice attains a depth of several feet, at least as much as four to seven. Pieces of floating ice of much greater thickness are sometimes seen, but it is doubtful if they represent the ice formed by the freezing of undisturbed water. At any rate, the ice formed in winter is often broken up in summer into floating pieces called *floe-ice* (Fig. 156). Floe-ice is sometimes crowded together in *ice-packs*, the separate pieces being so jammed together that some of them are ended up and stand high above the water. If the ice-pack of one summer is still far enough north at the end of the warm season, it is frozen together, and its aggregate thickness, made up as it is of blocks of ice some of which are on edge and much of it piled into ridges, is far beyond that of the normal sea ice.

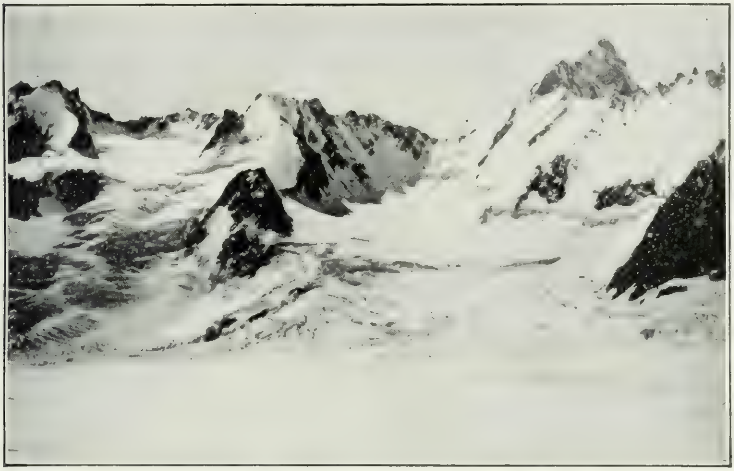


Fig. 157. — Snow-fields in the French Alps. (R. T. Chamberlin.)

Snow-fields. Over the larger part of the land, the snow of winter does not endure through the summer; but in cold regions where the fall of snow is heavy, some of it remains unmelted from year to year, and constitutes perennial snow-fields.

High mountains and the lands of high latitudes are the common habitats of snow-fields. In North America there are numerous small snow-fields in the western mountains, from Mexico on the south to Alaska on the north, their number and size increasing in the latter direction. In the United States there are few snow-fields south of the parallel of 36° 30′, and these are small, but farther north, and particularly in Alberta and British Columbia, the snow-fields of the western mountains attain considerable size. Snow-fields comparable to the latter occur in the higher mountains of Europe and Asia, while in South America there are snow-fields of small size even near the equator, and farther south, in the Chilean Andes, there are some of much greater dimensions. Small snow-fields occur on the highest peaks of tropical Africa, and larger ones in the mountains of New Zealand. For reasons which will appear later, a large part of every large snow-field is really ice.

Besides these fields of snow in mountain regions, there are fields of much greater extent in polar regions. The greater part of Greenland is covered with a single field of ice and snow, the size of which is estimated at 600,000 to 700,000 square miles, an area 400 to 600 times as large as the snow- and ice-covered area of Switzerland. In Antarctica there is a still larger field — the largest of the earth. Its area is only approximately known, but such data as are at hand indicate that it may have an extent ten times as great as that of Greenland.

The one condition necessary for a snow-field is an excess of snow-fall over snow-waste. The lower edge of a snow-field, the *snow-line*, is dependent chiefly on temperature and snow-fall. In general the snow-line does not depart much from the summer isotherm of 32° F., though it may be above this isotherm where the snow-fall is light. That the snow-line is not a function of temperature only, is shown by its position in various places. Thus in the equatorial portion of the Andes, the snow-line has an altitude of about 15,000 feet on the east side of the mountains, where the precipitation is heaviest, and of about 18,000 feet on the west side, where it is lighter. For the same reason the snow-line in the Himalayas is lower on the warmer, south side than on the cooler, north side. Though temperature and snow-fall are the most important factors controlling the position of the snow-line, humidity and the movements of the air are of some importance, since both affect the rate of evaporation of snow and ice.

Passage of Snow into Névé and Ice. The snow does not lie on the surface long before it undergoes obvious change.

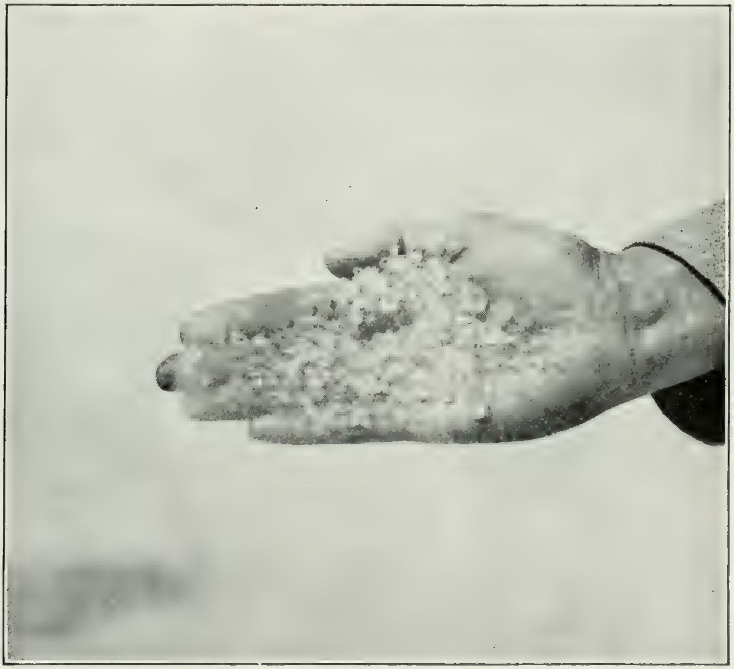


Fig. 158. — Ice granules in Sperry Glacier, Mont. (MacClintock.)

The light flakes are transformed into granules, and the snow becomes "coarse-grained." The smaller grains evaporate and the moisture condenses on the larger ones so that the larger ones grow at the expense of the smaller. The granular character, so pronounced in the last banks of snow in the spring, is even more distinct in perennial snow-fields. This granular snow is called *névé.* Where the thickness of the snow is considerable, the névé becomes more compact below, and finally passes into granular ice. Ice is found in most snow-fields at no great depth from the surface. Such ice is in a sense stratified, for it is made from successive falls of snow which to some extent retain their individuality.

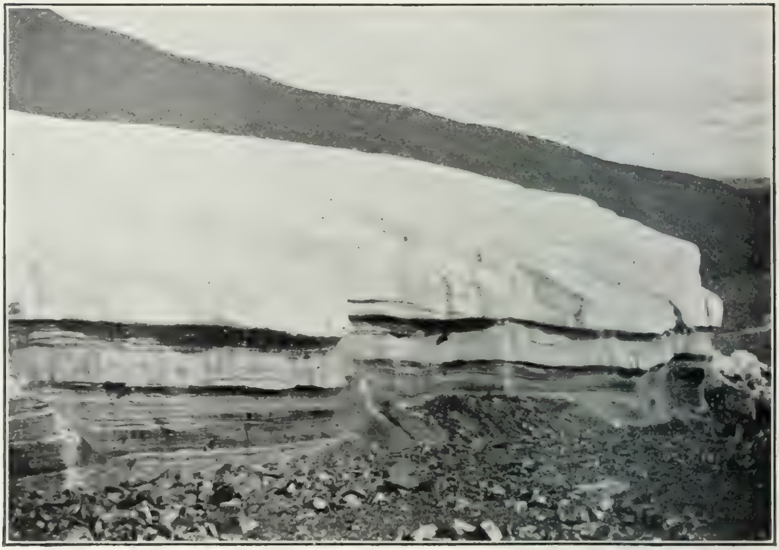


Fig. 159. — End of Bryant Glacier, a high-latitude glacier of North Greenland.

GLACIERS

When the body of snow and ice in a snow-field becomes sufficiently great, the force of gravity causes the mass to spread slowly outward or downward from its place of accumulation. This slow motion of the ice in an ice-field is *glacier motion*, and the ice thus moving is *glacier ice*.

Types. The different shapes of glaciers have given rise to different names. If the surface on which the ice-field develops is fairly horizontal, the ice will move outward in all directions, and ice spreading in all directions from a center is an *ice-cap*. The glacier covering the larger part of Greenland is a good example. The glaciers on some of the flat-topped peninsular promontories of the same island are good examples of small ice-caps. If ice-caps cover a large part of a continent, as some of those of the past have done, they are called *continental glaciers.*

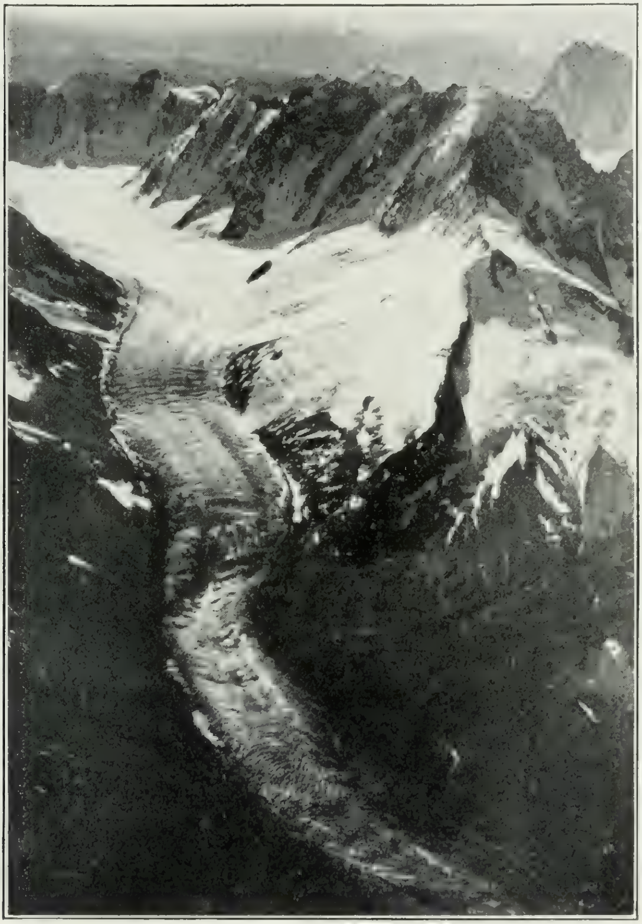


Fig. 160. — Glacier d'Argentiere, French Alps. (Swiss Air Serv.)

Where ice-caps lie on plateaus whose borders are trenched by valleys, ice-tongues from the edge of the ice-caps in places extend down into the valleys, and constitute one type of *valley glacier*. A second and more familiar type of valley glacier occupies mountain valleys, and is the offspring of mountain snow-fields. The former type, confined chiefly to high latitudes, is the *polar* or *high-latitude glacier* (Fig. 159); the latter is the *alpine glacier* (Figs. 160 and 161).

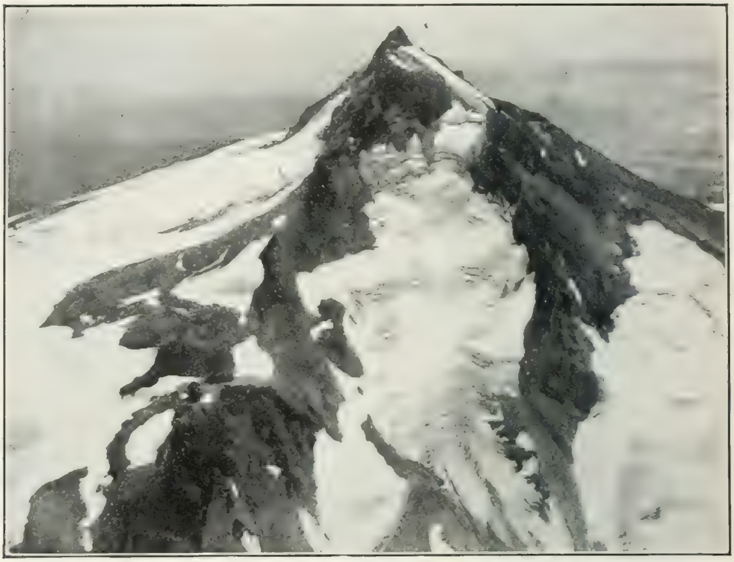
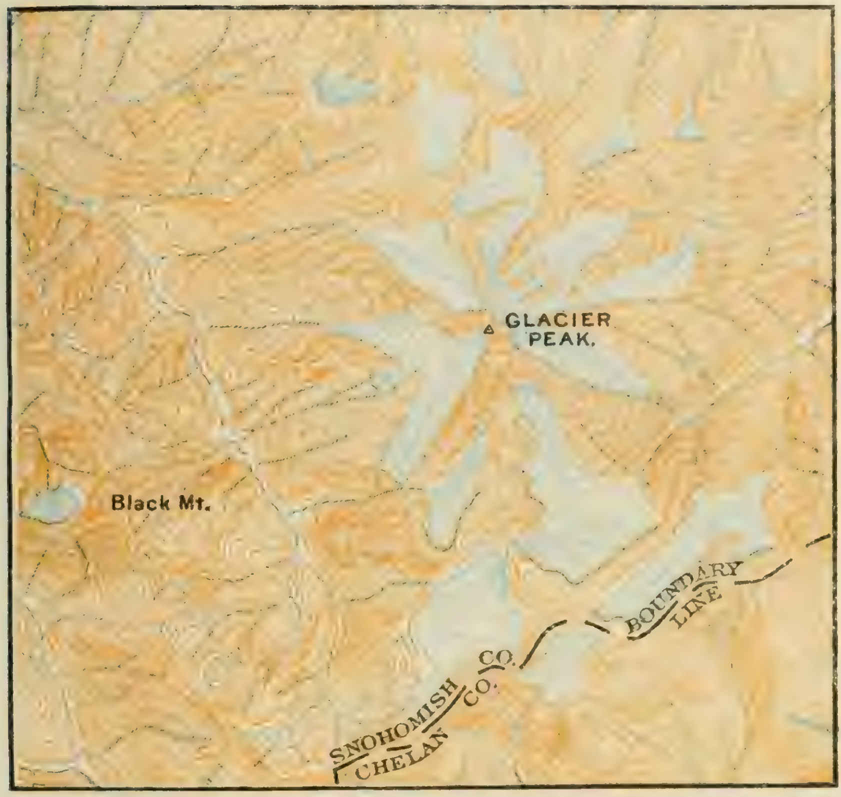


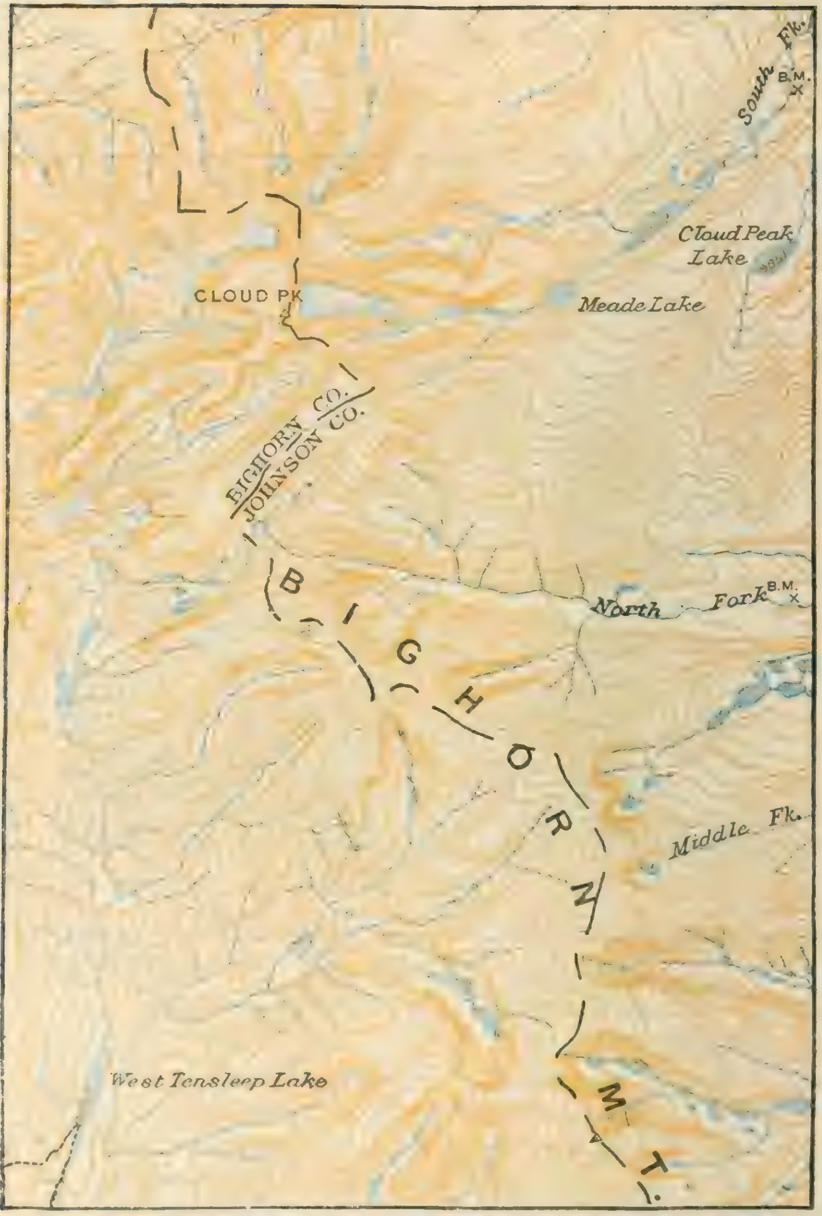
Fig. 161. — Valley glacier on Mt. Jefferson. (U. S. Army Air Serv.)

The end and side slopes of high-latitude glaciers are, as a rule, much steeper than those of alpine glaciers.

If a valley glacier descends through its valley to a plain beyond, its end spreads. The deploying ends of adjacent glaciers in places merge to form a *piedmont glacier* (Fig. 162). Piedmont glaciers are confined to high latitudes. In some cases the snow-field that gives rise to a glacier is restricted to a relatively small depression in the side of a mountain, or in the escarpment of a plateau. In such cases the snow-field and glacier are hardly distinguishable, and the latter descends but little below the snow-line. Such a glacier nestled in the face of a cliff has been called a *cliff glacier* (Fig. 163).


GLACIER PEAK.
Black Mt.
SNOHOMISH CO.
CHELAN CO.
BOUNDARY LINE

Glaciers on Glacier Peak, Washington. Scale, about 2 miles per inch. (Glacier Peak Sheet, U. S. Geol. Surv.)


South Fk.
B.M.
Cloud Peak Lake
Meade Lake
CLOUD PK
BIGHORN CO.
JOHNSON CO.
BIGHORN MT.
North Fork
B.M.
Middle Fk.
West Tensleep Lake

A portion of the Bighorn Mountains, showing glaciated valleys, the heads of which are in many cases cirques. Scale, about 2 miles per inch. (Cloud Peak, Wyo., Sheet, U. S. Geol. Surv.)

Cliff glaciers, often as wide as long, and always small, grade into valley glaciers. Occasionally the end of a valley glacier, or the edge of an ice-sheet reaches a precipitous cliff, and the end or edge of the ice breaks off and accumulates below like talus. The fragments of ice may then become a coherent mass by reunion (regelation), and the whole may resume motion. Such a glacier is called a *reconstructed glacier*. There are many examples of this along the precipitous cliffs of the Greenland coast.

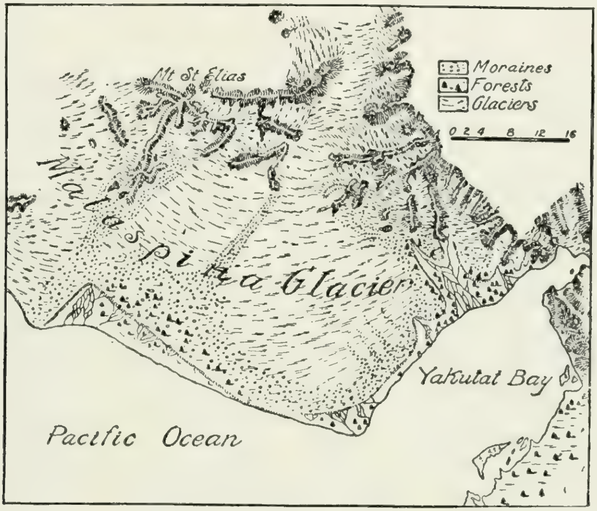


Fig. 162. — Malaspina Glacier, a piedmont glacier in Alaska. (After Russell.)

ICE-CAPS

Of the foregoing types of glaciers, ice-caps far exceed all others both in size and geologic importance. At the present time, however, only two, those of Greenland and Antarctica, are large enough to be called continental glaciers. But in the geologic past great ice-caps have occupied parts of all the continents.

The Greenland ice-cap occupies most of the island, leaving only a fringe of unglaciated land around the margins. The ice rises from a thin edge to an altitude of 8000 or 9000 feet in the interior. The thickness of the ice, however, is not known, but some thousands of feet would probably not be too large an estimate. Its movement, differing from place to place and from time to time, averages not more than a foot or so per week. But where the tongues of ice project from the main mass into the valleys around the border the speed may be many times this amount. The surface of the ice-cap is a nearly featureless snowy plain above which rise a few ice-free peaks such as Mount Forel (9050 feet elevation), near the center of South Greenland.

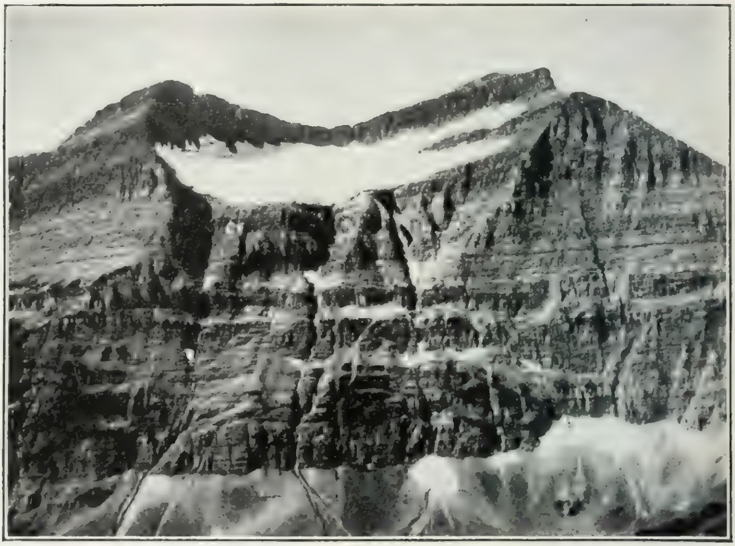


Fig. 163. — Cliff glacier on Piegan Mountain, Glacier National Park, Mont. (R. T. Chamberlin.)

The Antarctic ice-cap, while exceeding that of Greenland in area and possibly in thickness, presents an interesting case of imperfectly consolidated ice. Probably the exceedingly low temperature of the polar region partially prevents compacting and solidification common to less rigorous climates.

THE VALLEY GLACIER

Since valley glaciers are the most numerous and have been best studied, the general phenomena of glaciers will be discussed with primary reference to them. A valley glacier in general conforms to the shape of the valley in which it lies (Fig. 164). If the valley is wide the glacier is wide; if crooked the ice curves with the irregularities of the walls.

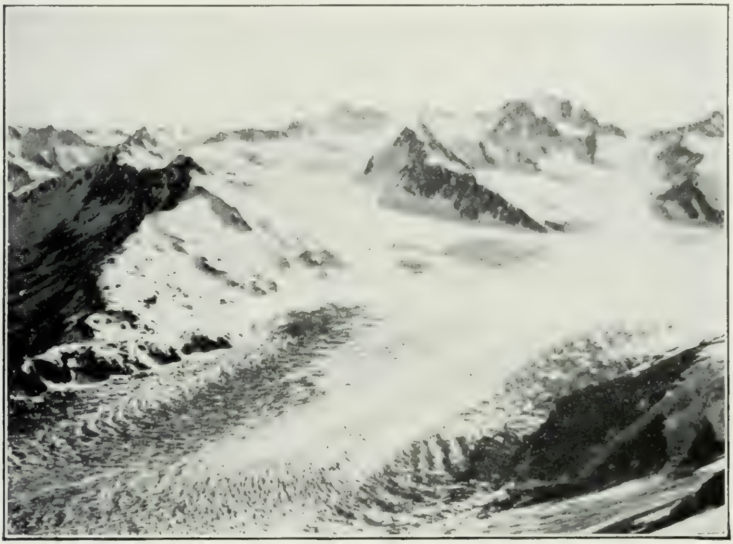


Fig. 164. — Childs Glacier, Alaska. A huge valley glacier fed by many tributaries. (R. T. Chamberlin.)

It bears some resemblance to the non-technical description "a river of ice"; but a glacier so differs from a river that the description, if seriously considered, would be misleading.

Its Surface. The upper reaches of a valley glacier are covered with snow the year round, giving a rather smooth or undulatory surface, broken here and there by a few large cracks (*crevasses*) (Fig. 165). Toward their lower ends most glaciers commonly become free from snow during the summer and in most cases present a much rougher appearance, due chiefly to the large number of crevasses which open in the more rapidly moving portions of the glacier (Fig. 166).

This cracking is due to tension in the ice developed especially where the brittle ice rides over an uneven bed and where some parts of the glacier move more rapidly than others. Where there is an abrupt increase of gradient in the bed of the glacier, an ice-fall results, and the ice is greatly riven. Crevasses in the middle of a glacier are commonly at right angles to the axis of the glacier. Those along the sides of a valley glacier point upstream from the ice-edge toward the middle of the glacier. This is because the middle of the glacier, moving faster than the sides, develops tension and the crevasses open at right angles to the tension. Longitudinal crevasses also form when the foot of a glacier spreads (Fig. 167). This breaking of the ice as it moves is one of the striking differences between a glacier and a river, a stream of lava, or any other liquid substance.

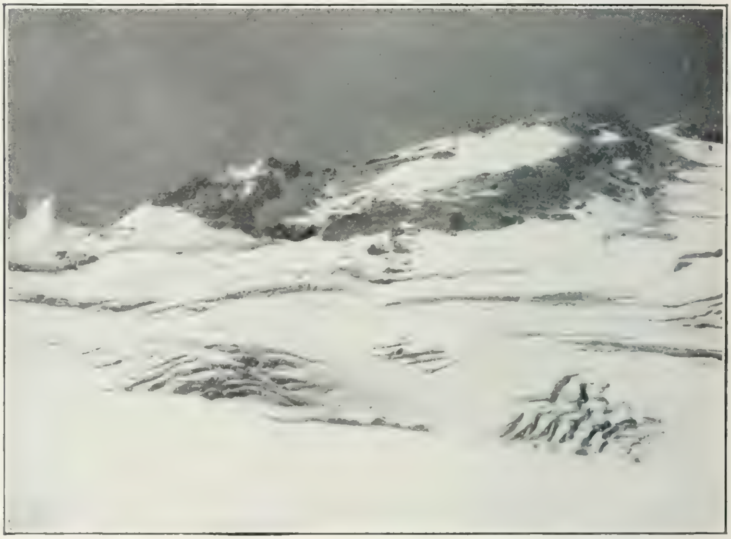


Fig. 165. — Crevasses in the Asulkan Glacier. British Columbia. (R. T. Chamberlin.)

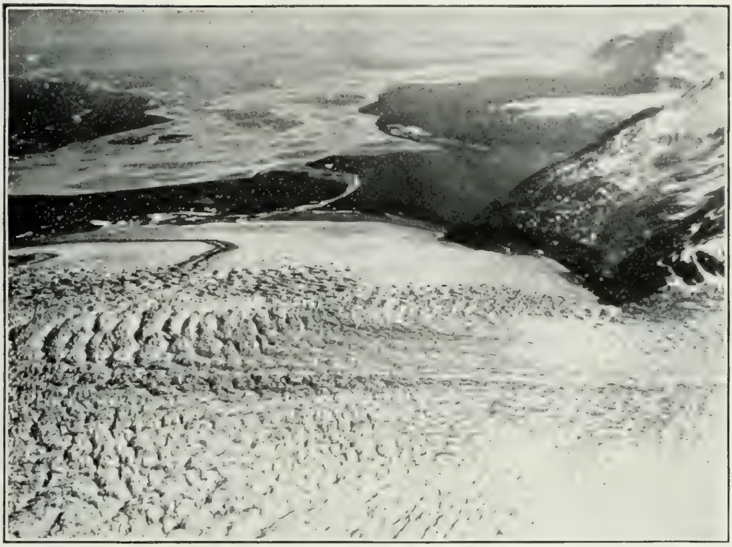


Fig. 166. — The crevassing of Childs Glacier, Alaska. The glacier is nearly three miles wide. Note the braided Copper River beyond. (R. T. Chamberlin.)

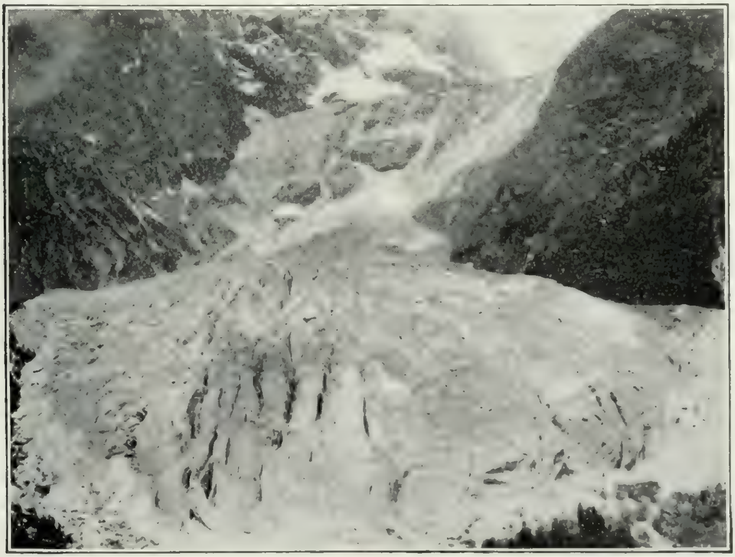


Fig. 167. — Glacier spreading near its terminus, developing longitudinal crevasses. Brenva Glacier, Italian Alps. (R. T. Chamberlin.)

In the forward movement of the ice many of the crevasses close; but in many cases while they are open, warm air and the heat of the sun so melt the ice and widen them, especially the upper parts, that when the edges come together they do not fit closely. The result is a very uneven surface.

The lower ends of many glaciers extend below the snow-line and into the zone of melting temperature during the summer months. Surface water gathers in small pools or drains off, either to the edge of the glacier or to crevasses into which it plunges. While flowing on the surface, such water develops small channels (valleys) in the ice (Fig. 168). Where *superglacial* streams plunge into crevasses they wear, or melt out, well-like holes (moulins) in the ice, and if whirling motion is acquired may even develop pot-holes in the rock beneath.

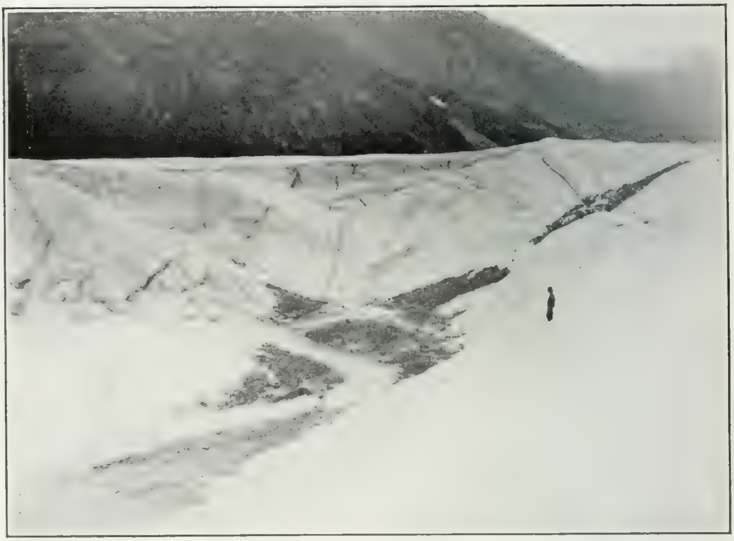


Fig. 168. — Superglacial stream in small valley on the Kennecott Glacier, Alaska. Debris has collected in the valley. (R. T. Chamberlin.)

Surface Moraines. The surface of most glaciers is littered with rock debris, which is commonly disposed in the form of belts or *moraines* (Fig. 169). The surface moraines may be *lateral, medial,* or *terminal.* A *lateral moraine* is any considerable accumulation of debris in a belt on the side of a glacier (Fig. 170). A *medial moraine* is a similar accumulation at some distance from the margins, but not necessarily in or even near the middle. There may be several medial moraines on one glacier, in which case some of them may be far from the middle (Fig. 171). On valley glaciers which have sufficiently gentle terminal slopes a *surface terminal moraine* may develop which, in some cases, connects two lateral moraines, making a loop roughly concentric with the end of the glacier.

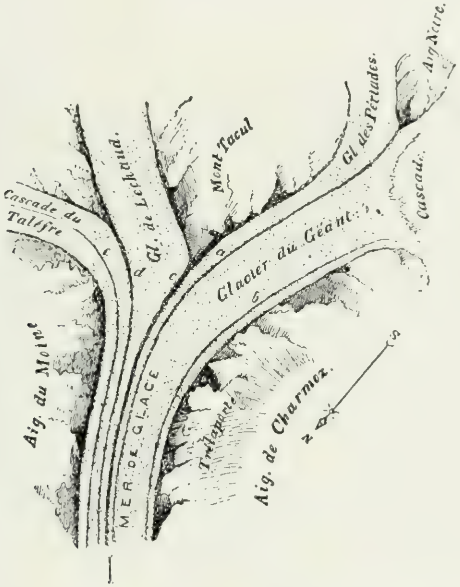

Aig. Noire.
Gl. des Périades.
Cascade.
Mont Tacul
Gl. de Lechaud.
Cascade du Taléfre
Glacier du Géant
Aig. du Moine
MER DE GLACE
Trélaporte
Aig. de Charmoz.
S
N
a
b
c
d
e

Fig. 169. — Figure showing the union of glaciers and the development of medial moraines by the union; also the position of lateral moraines. (After Tyndall.)

Besides the surface moraines, there may be scattered boulders and bits of rock of various sizes on the ice, and, in addition to the coarse material, there is considerable dust present.

Relief Due to Surface Debris. The debris on the ice, by influencing melting beneath and about itself, produces surface irregularities. The rock debris absorbs heat more readily than the ice. A thin piece of stone lying on the ice is warmed through by the sun's rays, and, melting the ice beneath, sinks, just as a piece of black cloth on snow will sink because of the increased melting beneath it. Similarly dust, because it absorbs heat readily due to its dark color, and because it is thin enough to be heated through, melts the underlying ice to form small depressions called *dust-wells*. Though a good absorber of heat, rock is a poor conductor, and so the lower surface of a thick mass of stone is not warmed notably, and the ice beneath, being protected from the sun, is melted more slowly than that around it. The result is that the boulder presently stands on a protuberance of ice (Fig. 172). When its pedestal becomes high, the oblique rays of the sun and the warm air surrounding it cause it to

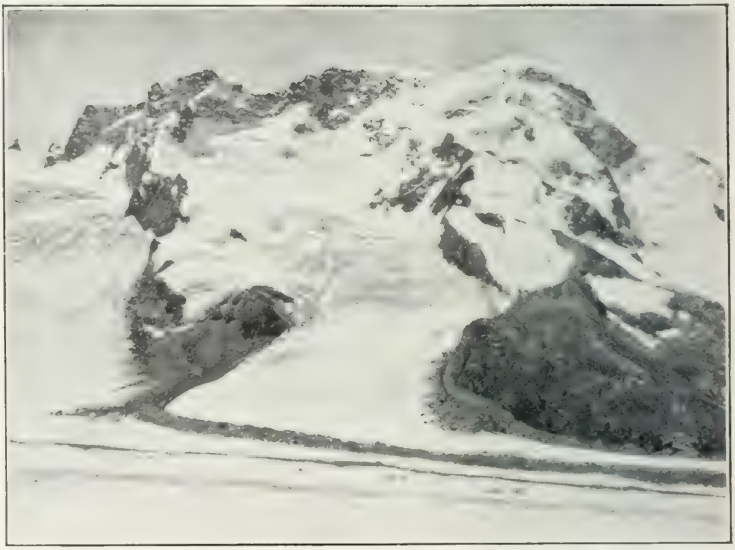


Fig. 170. — Breithorn Glacier, Swiss Alps. Shows source of material which makes lateral moraine. Lateral moraine of this glacier becomes a medial moraine on the combined glacier below the junction. (R. T. Chamberlin.)

waste away, and the capping boulder falls. Similarly, unusually dense patches of earthy material may so protect the ice beneath from melting as to give rise to conspicuous *ice-cones* thinly mantled with a couple of inches of dirt (Fig. 173).

The same principle applies to the moraines. A surface moraine usually protects the ice beneath from melting, and causes the development of a ridge of ice beneath itself, so that what appears to be a ridge of debris is in reality an ice-ridge only veneered with rock material. As the ice on either side is lowered by ablation, the moraine material tends to slide down on either hand. So far does this spreading go that in some cases the lower end of a glacier is completely covered with the debris which has spread from the medial and lateral moraines.

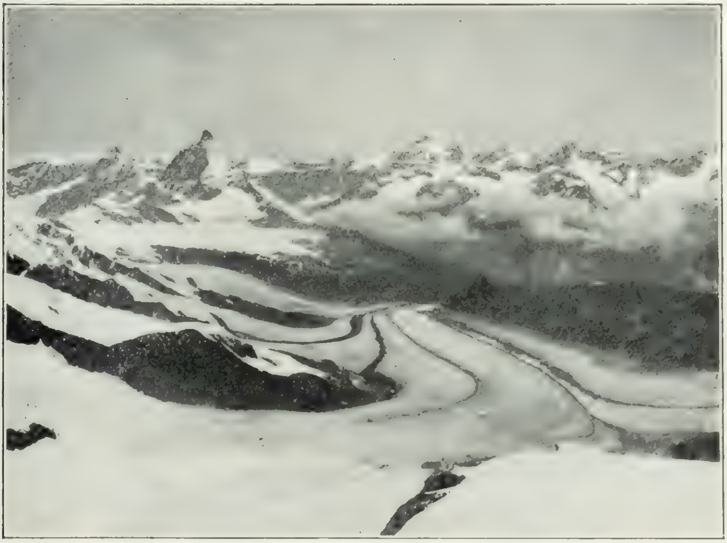


Fig. 171. — Gorner Glacier with its tributaries, the Grenz, Schwarze, Breithorn and Theodule glaciers. Many medial moraines. From summit of Monte Rosa, Switzerland. (R. T. Chamberlin.)

Debris Below the Surface. The lower part of a glacier, as well as the upper, carries rock debris. This debris is sometimes and in some places so abundant, especially near the ends and edges of the ice, that it is difficult to locate the bottom of the glacier, for between the moving ice, which is full of debris, and the stationary debris, which is full of ice, there seems to be, in many cases, a nearly complete gradation. The debris in the lower part of arctic glaciers, and to some extent of others, is often disposed in thin sheets sandwiched between layers of clean ice. Debris also occurs to some extent in the ice far above its base, in sheets, in bunches, or scattered (Figs. 174 and 175).

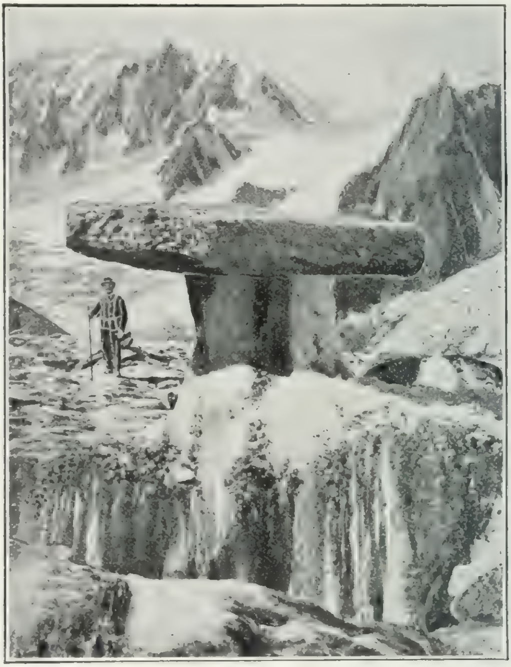


Fig. 172. — A glacial table due to the protection of the ice beneath the flat stone from the rays of the sun. Talèfre Glacier.

Drainage. Surface water either finds its way to the edge or end of a glacier, drops below the surface through crevasses, or sinks through the leaky glacial ice. Water below the surface may freeze, but in thin glaciers and near the lower end of most glaciers it is likely to collect in more or less definite streams either at the bottom of the glacier, *subglacial streams*, or within the body of ice, *englacial streams*. In alpine glaciers, the waters beneath the ice often unite in a common stream in the axis of the valley, and hollow out a tunnel. As an example, the Rhone River is already a considerable stream where it issues from beneath the Rhone Glacier.

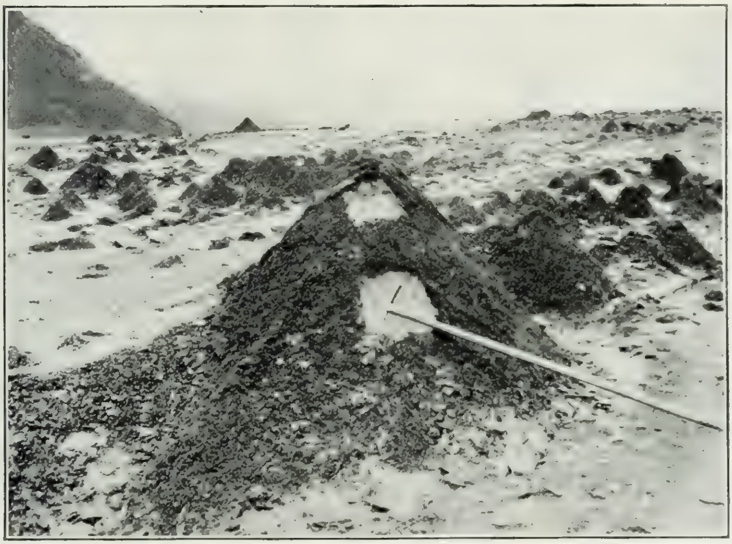


Fig. 173. — Ice-cones on Valdez Glacier, Alaska. A veneer of debris prevents ready melting of the ice beneath. (R. T. Chamberlin.)

At the end of a valley glacier, all waters, whether they have been superglacial, englacial, or subglacial, unite to bear away the silt, sand, gravel, and even small boulders set free from the ice, and to spread them in belts fringing the borders of the ice, or in *trains* stretching down the valleys, *valley trains*.

Movement. The ends of glaciers are sometimes observed to remain in a certain place year after year. Observations during the same time show that melting of the ice of the end is constantly going on. It thus becomes clear that the edge remains stationary because the *ice of the glacier is moving forward at the same rate* as that at which it is being melted at the end. Other observations show that the ends of glaciers at times advance farther down the valleys, possibly overturning trees or other familiar objects. This evidently means that the ice of these glaciers is then moving faster than it is being melted at the end. At some other times or in still other cases, the ends are observed to retreat, showing that melting exceeds the forward motion of the ice, it being obvious that the ice of the glacier does not move up the valley when the end retreats.

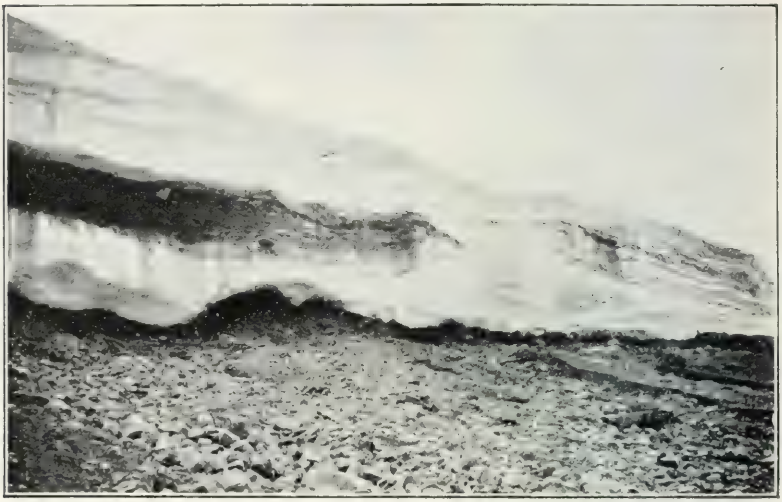


Fig. 174. — Side view of end of a glacier. Southeast side of McCormick Bay, North Greenland. Shows structure of ice as well as position of debris.

Rate of Movement. By driving stakes in the ice, or using other markers, it is found that the middle of a valley glacier moves faster than the sides. The movement varies with different glaciers and at different times for the same glacier. In general it ranges from several inches up to several feet per day, though some of the glaciers of Greenland have been found to move 50 or 60 feet, and one even 100 feet, per day in the summer time; but these high rates have been observed only where the ice of a large inland area crowds down into a comparatively narrow fiord, and debouches into the sea, and then only in summer.

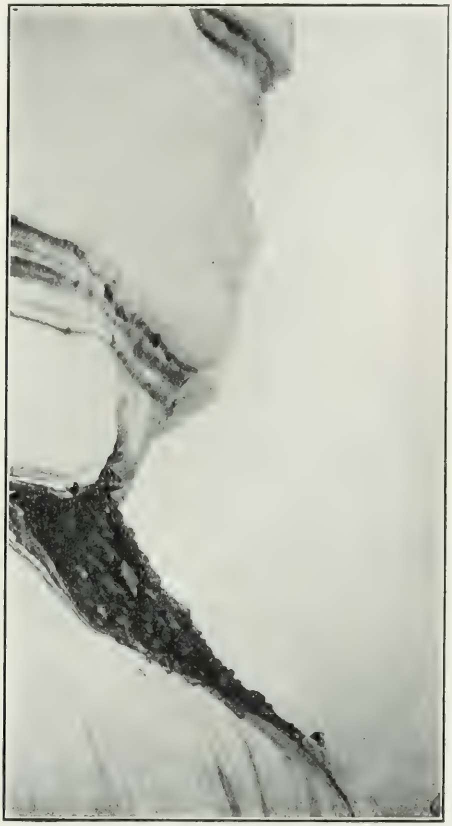


Fig. 175. — A part of the vertical side of a North Greenland glacier. The vertical or even overhanging faces are sometimes more than 100 feet high.

Conditions Affecting Rate of Movement. The rate of glacier movement appears to depend on (1) the depth of the moving ice, (2) the slope of the surface over which it moves, (3) the slope of the upper surface of the ice, (4) the topography of its bed, (5) temperature, and (6) the amount of water in the ice. Great thickness, steep slopes, smoothness of bed, a high (for ice) temperature, and abundance of water favor rapid movement. Since some of these conditions, notably temperature and amount of water, vary with the season, the rate of movement of a glacier varies during the

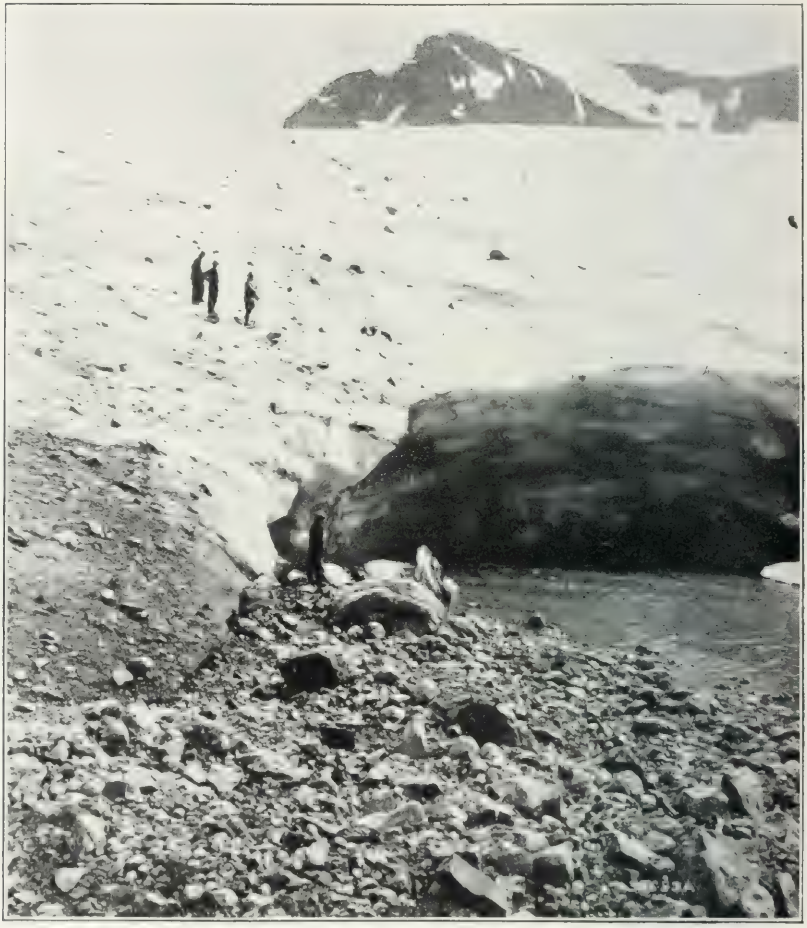


Fig. 176. — Stream issuing from a glacier. Lyman Glacier, Chelan Forest, Wash. (Courtesy U. S. For. Serv.)

year. Other conditions, especially the first of those mentioned above, vary through long periods of time, and cause corresponding variations in the rate of movement.

A declining upper surface is essential to glacier movement. To be sure, there are short stretches where there is no decline; indeed, there are occasional places where the upper surface slopes backward, as where the ice is pushed up over a swell in its bed; but such cases are local exceptions, and do not militate against the general truth of the statement that the upper surface of a glacier slopes in the direction of motion. A declining *lower* surface is less necessary. In the case of valley glaciers, the bed does, as a rule, decline in the direction of motion; but that there are exceptions is shown by the deep basins in the rock which such glaciers leave behind them when they retreat. In the great continental glaciers of not far remote geologic time, the ice frequently moved up slope for scores, and even hundreds of miles; but in all such cases, the prevailing slope of the upper surface must have been *down* in the direction of movement.

NATURE OF MOVEMENT

The apparent similarity between rivers and glaciers led early observers to consider glacier-motion as the flow of a rather stiff, viscous *liquid.* But even casual consideration makes it evident that, since cracking open is not a characteristic of flowing substances, a cracked and crevassed glacier is not flowing like a liquid. Moreover, glaciers hold pieces of rock so firmly as to gouge and scratch other rocks. This could not be accomplished by a liquid substance. Glacier-ice is a solid crystalline substance, its molecules oriented in fixed attitudes, and it must move as a solid.

Glacier movement is a complex process, but some of the important elements seem to be: (1) exchange of individual molecules between granules *(idiomolecular exchange)*, (2) rotation of granules, (3) melting and re-freezing, involving transfer of considerable amounts of water, and (4) shearing or sliding within the mass of the glacier. While these elements are inextricably bound together in operation, it is helpful to discuss them separately.

1. Idiomolecular Exchange. As has already been stated, the passage from snowflakes to glacial ice is a process of granulation. The larger granules grow at the expense of the smaller ones. This is doubtless due to the fact that the sharper curvature of the smaller grains gives less secure attachment for the molecules than the gentler curvature of the larger grains (i.e., the smaller ones have a higher vapor pressure). The consequent constant interchange of molecules between grains finds equilibrium in the growth of the larger ones. This interchange and growth tend to relieve strains within the ice by the passage of molecules in the direction of least resistance and thus express themselves in a downward and outward movement of the body of ice.

Matter tends to adapt itself to the intensity conditions of heat and pressure. These express themselves in the case of ice by a movement, molecule by molecule, of the matter from a place of higher intensity to one of lesser intensity. Pressure at one point of a granule will cause movement of the molecules away from that point and their attachment at another point of less pressure. The movement may pass through the crystalline structure of the granule, or around it, or both. The total result of these innumerable idiomolecular transfers from places of greater to those of lesser pressures throughout the whole of the glacier results in glacial motion of the mass.

2. Rotation of Granules. Compression of ice generates heat, and if the compression is great enough this heat will melt the ice. In addition, pressure lowers the melting point, which facilitates melting. Glacier-ice is not compact and dense, but, as we have seen, is composed of crystalline granules with considerable pore-space among them. It is evident that each granule is in contact with its neighbors at only a few places. Pressure consequently is effective in producing melting only at those points of contact. Furthermore, it follows that the water generated by such melting moves only a little way before the pressure is so lessened that it re-freezes. We may picture, then, momentary liquefaction at innumerable points within the glacier. But during such momentary liquefaction, adjustments of a slipping, compacting, and rotating nature take place between the granules, resulting in motion of the whole mass down the valley (Fig. 177).

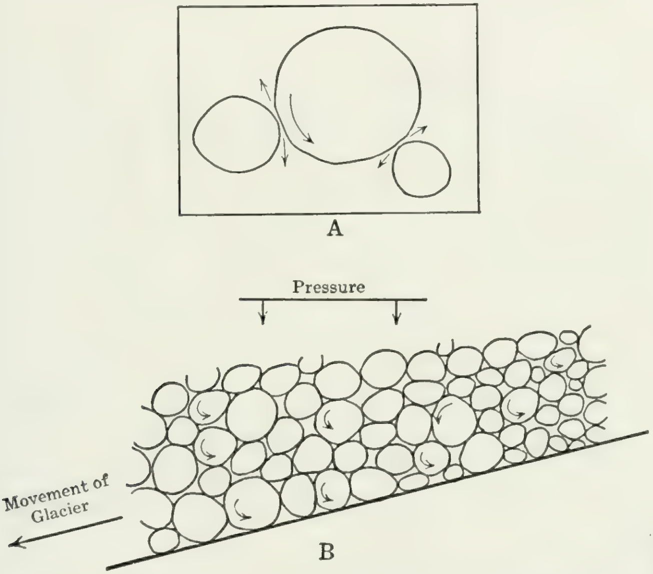


Fig. 177. — A. Melting at points of contact allows granules to rotate.
B. Rotation of granules producing glacier-motion.

3. Water Transfer. Water expands on freezing. During the warm months the heat of the sun is constantly melting the surface of glaciers. Part of this water, particularly near the end or sides of the glacier, flows away, but part of it flows or seeps down into the body of the glacier where the temperature is below freezing. When this water expands upon freezing, the thrust outward in all directions is most effective *down* the valley, where the resistance is least. The accumulation of this thrust tends to push the mass of the glacier bodily down the valley. Furthermore, the water in transit from the place of melting to that of freezing not only moves down into the ice but also down the valley. This results in the transfer of the substance of the glacier towards its lower end, but it does not contribute much to the powerful motion which scours the glacier-bed.

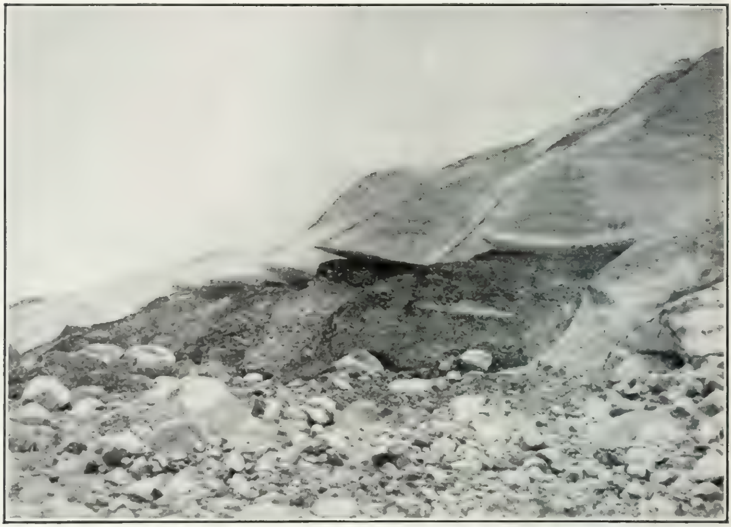


Fig. 178. — Shearing of upper portion of a glacier over the less movable portion below. Mendenhall Glacier, Alaska. (R. T. Chamberlin.)

4. **Shearing.** Down-valley from the snow-fields, where normally the thrusts are greatest, the granules fewest, and their interlocking most intimate, shearing takes place within the ice itself. Because the basal portion of a glacier encounters more resistance to forward motion than that above, the overlying portion of the ice commonly slides or shears over that beneath along many distinct shearing planes (Fig. 178). This takes place not only between the basal

and the overlying ice, but also between the terminal portion and that farther up-valley.

It is believed, in summary, that glacier-motion is not that of a viscous body, but essentially that of a rigid, crystalline solid, augmented by the water contained in the body of the glacier.

THE WORK OF GLACIERS

EROSION AND TRANSPORTATION

Valley glaciers abrade the valleys through which they pass, and carry forward the material which they remove from the surface, wearing, grinding, and ultimately depositing it. Like other agents of gradation, therefore, their work includes erosion, transportation, and deposition.

Getting Load. If the snow-field accumulates on a rough surface covered with abundant rock debris, the developing glacier already has a basal load when it begins to move, for the snow covers, surrounds, and includes such loose blocks of rock as project above the general surface, and envelops the projecting points of rock within its field. When the ice begins to move, it carries forward this loose rock in its bottom, and tears off the weak points of the enveloped rock-projections. It may also drag along some of the earthy matter of the surface on which the ice forms, and to which it is frozen. In addition to the *basal* and *subglacial* load which the glacier has at the outset, there may be, in the case of valley glaciers, surface debris which has fallen on the snow and ice from cliffs above. This is likely to be the case where steep cliffs tower above the snow-fields. If debris descending to the glacier in this way is unburied, it is *superglacial*, but if it has been buried by subsequent falls of snow, it is *englacial*.

Once in movement, the ice not only moves the originally incorporated debris but gathers more load. This it acquires partly by the rasping effect of its rock-shod bottom, and partly by its power of *plucking* off, or quarrying out, considerable blocks of rock from the valley sides and bottom. This plucking process is most effective where the ice passes over small cliffs of jointed rock, but it is not confined to such situations. The steep bed of a valley glacier in many cases is worn more by plucking than by rasping. The advancing ice gets some material, too, especially loose debris along the sides, by freezing to it. Superglacial material may be acquired during movement, as well as before it, by the fall of debris from the cliffs, or by the descent of avalanches.

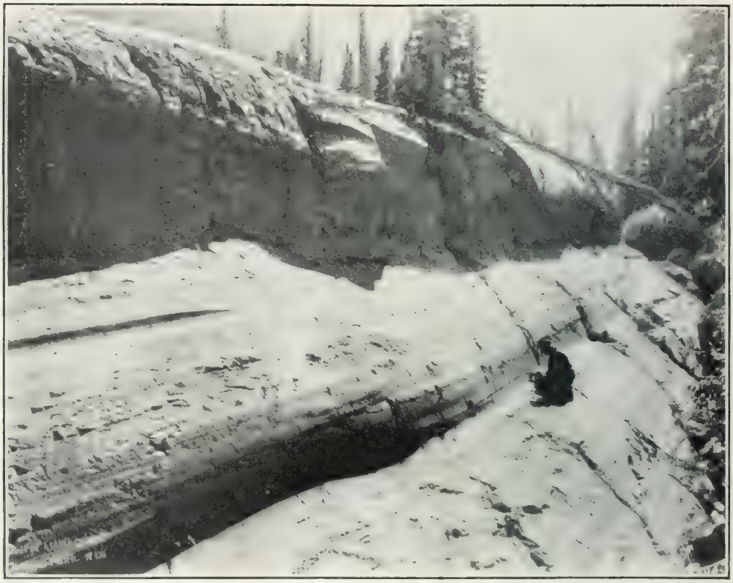


Fig. 179. — Groves, fluting, and striae. Glacier Creek, Mont. (Courtesy U. S. For. Serv.)

Conditions Influencing Rate of Erosion. (1) A somewhat obstructive configuration of the surface over which the ice passes is necessary for effective erosion. Ice wears a flat surface relatively little since there is little for it to get hold of. In places ice may override such a surface, burying its soil and even more or less of its herbaceous vegetation.

Erosion is probably at its maximum, so far as influenced by topography, where the surface is rough enough to offer notable catchment for the base of the ice, but not so rough as seriously to impede its motion. The amount of roughness favorable for the greatest erosion increases with increasing thickness of the ice.

Other conditions which influence erosion by ice are (2) the amount of loose debris on the land surface; (3) the resistance of the rock; (4) the slope of the land surface; (5) the thickness of the ice; (6) the rate of its movement, and (7) the abundance and character of debris which it has to work with. The effect of most of these conditions is evident, but the effect of the last is less simple. Clean ice passing over a smooth surface of solid rock would have little effect upon it; but a rock-shod glacier will abrade the same surface notably. The effect of this abrasion is shown in the grooves and scratches (*striae*) which the stones in the bottom of the ice inflict on the surface over which they pass

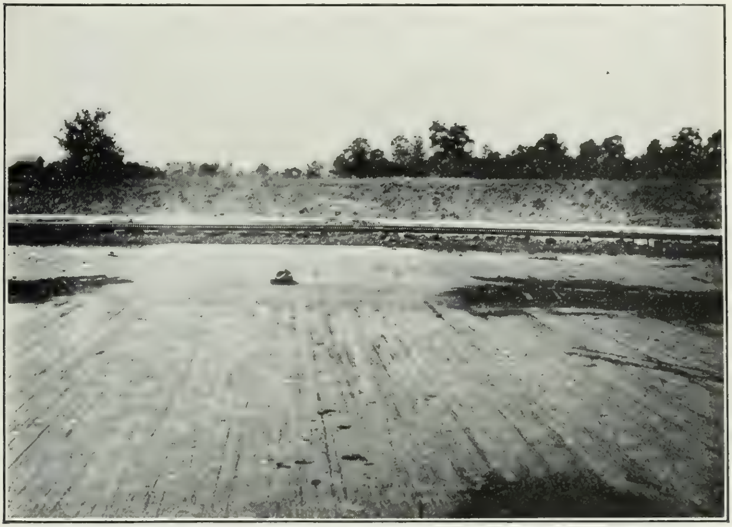


Fig. 180. — Striae on bed rock. (Stauffer.)

(Figs. 179 and 180). At the same time, the stones in the ice are themselves worn by abrasion with the bottom (Fig. 181). It does not follow, however, that erosion is greatest when there is most material in the bottom of the ice; for with increase of debris there may be decrease of motion, and decrease of motion seriously reduces the effectiveness of erosion. When any considerable thickness of ice at the bottom of a glacier is full of debris, the loaded portion may approach stagnation, while the cleaner ice above shears over it. A moderate, but not excessive load of debris is, therefore, favorable for great erosion. Something depends, too, on the character of the load. Coarse, hard, and angular debris is a more effective instrument of abrasion than fine, soft, or rounded material. So far as plucking is concerned, rate of motion is probably more important than load.

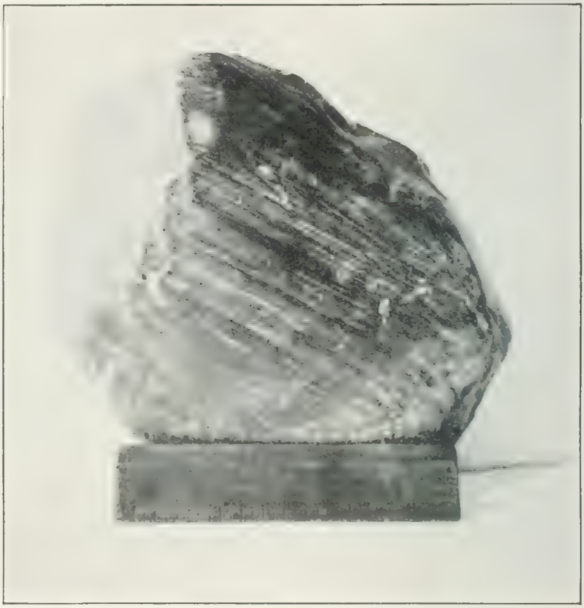


Fig. 181. — Stone striated by glacial wear.

In connection with the resistance of the rock, it should be noted that resistance is not simply a matter of hardness and softness. Rock which is affected by bedding planes, by cleavage planes, or joints is obviously more easily eroded than rocks which do not have these planes of weakness. On steep slopes and beneath the ice-field, plucking is probably more important, on the whole, than wear by the load.

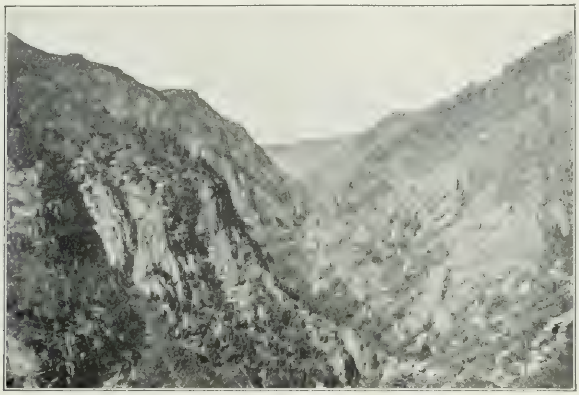


Fig. 182. — An unglaciated valley in the Wasatch Mountains. (Church.)

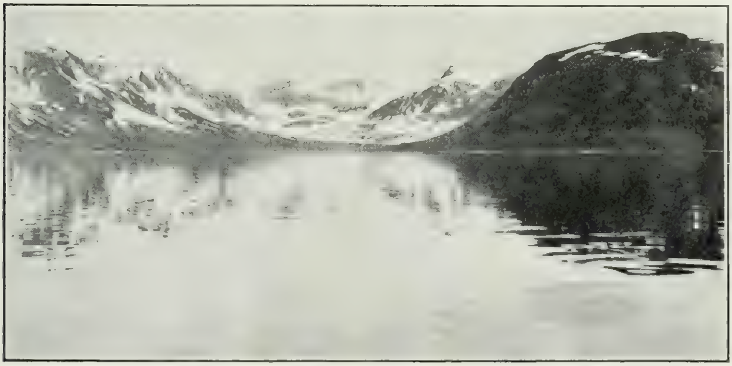


Fig. 183. — A U-shaped glaciated mountain valley on the coast of Norway. (R. T. Chamberlin.)

In summary it may be said that rapidly-moving ice of sufficient thickness to be working under goodly pressure, shod with a sufficient, but not excessive, quantity of hard rock material, passing over non-resistant formations possessing a topography of sufficient relief to offer some resistance, and yet too little to retard the progress of the ice seriously, will erode most effectively.

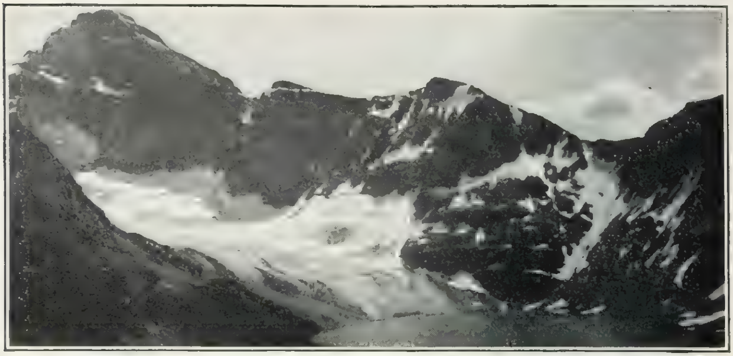


Fig. 184. — Small glacier in its cirque. British Columbia. (Porter.)

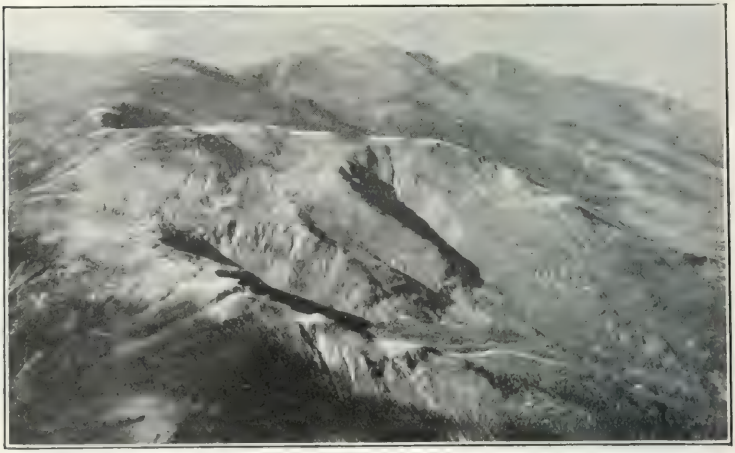


Fig. 185. — Cirques in the White Mountains, N. H. Mt. Washington, Tuckerman's and Huntington's ravines. (Courtesy U. S. Army Air Serv.)

Varied Nature of Glacial Debris. From its manner of erosion it will be seen that a glacier may be charged with various sorts of material. At its bottom there may be (1) boulders which the ice has picked up from the surface, or which it has broken off from projecting points of rock over which it has passed; (2) smaller rocks of the size of pebbles, cobbles, etc., either picked up by the ice from its bed, or broken off from larger masses; (3) the fine products (*rock-flour*) produced by the grinding of the rock-shod glacier on the rock-bed over which it passes, and similar products resulting from the rubbing of stones together along shearing planes; and (4) sand, clay, soil, vegetation, etc., derived from the surface overridden. Thus the materials which the ice carries are of all grades of coarseness and fineness, from huge boulders to fine clay. Whether one sort of material or another predominates depends primarily on the nature of the surface and materials overridden.

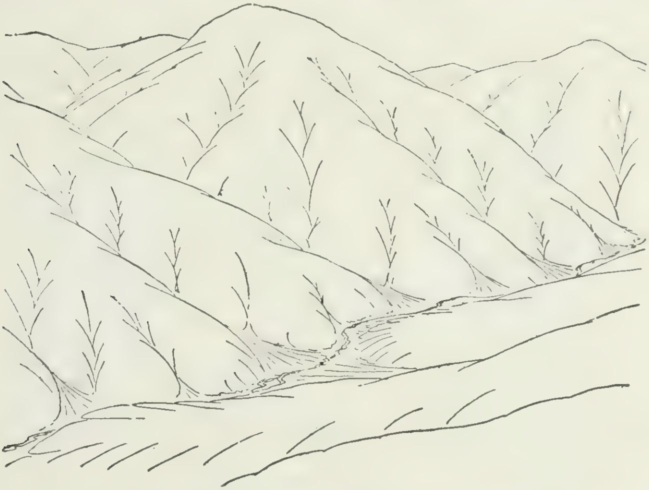


Fig. 186. — A normally eroded mountain area not affected by glaciation. (Davis)

Topographic Effects of Glacial Erosion. An alpine glacier widens, deepens, and smooths the valley through which it passes. A steep V-shaped valley is commonly made U-shaped by widening and rounding the bottom. Such widening is very noteworthy at the heads of many mountain

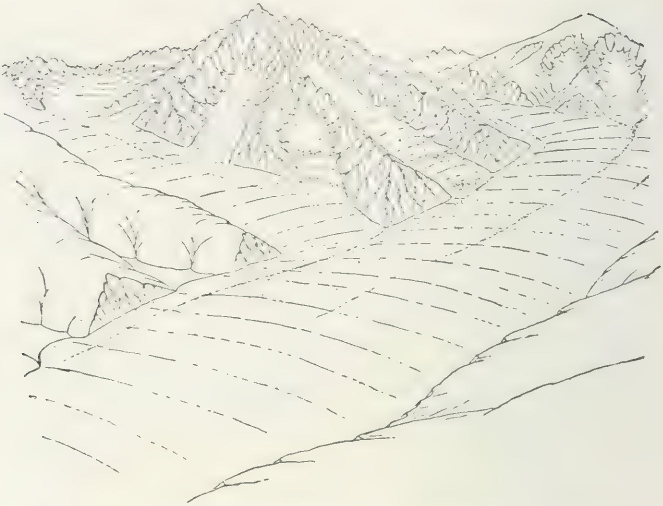


Fig. 187. — The same area shown in Fig. 186 affected by glaciers which still occupy its valleys. (Davis.)

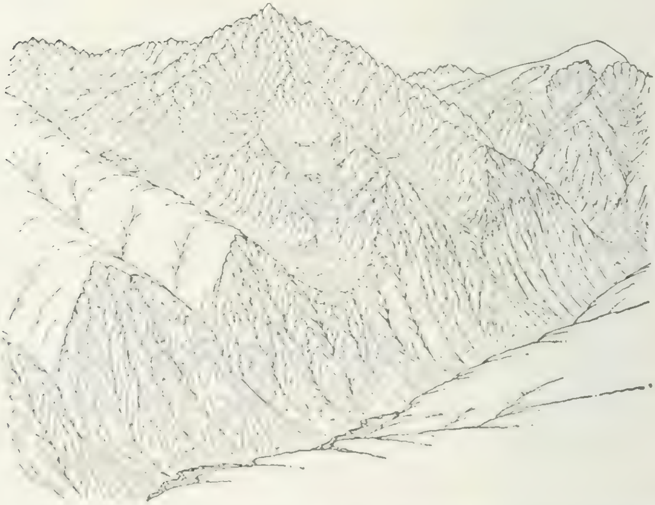


Fig. 188. — The same area shown in the two preceding figures after the ice has melted. (Davis.)

valleys, where steep-walled, amphitheatre-like valley-heads, *cirques*, are produced.

The deepening and widening of a valley by glacial erosion may throw its tributaries out of topographic adjustment. Thus, if a main valley is lowered 100 feet by glacial erosion while its tributary is not correspondingly deepened, the lower end of the latter will be 100 feet above the former when the ice disappears. Such a valley is called a *hanging valley* (Figs. 186, 187 and 188). In many rugged mountain districts which have been glaciated, tributary valleys

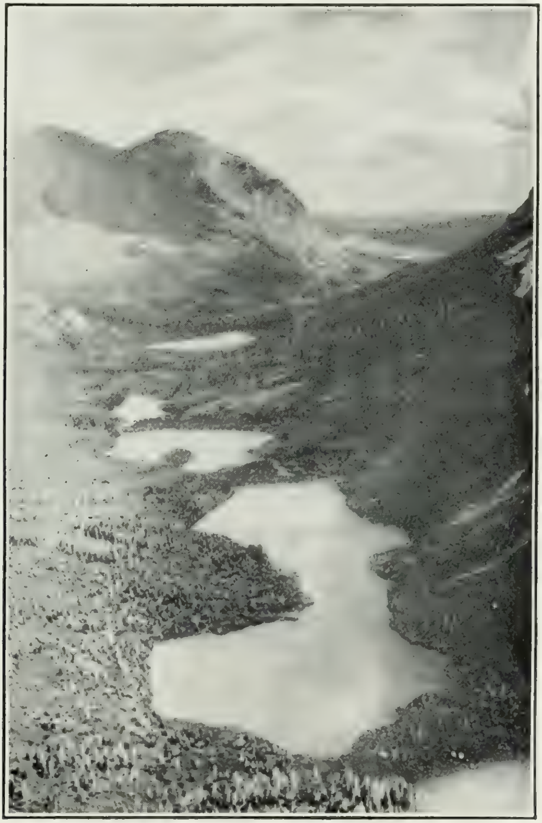


Fig. 189. — U-shaped valley and glacial lakes. Montana. (MacClintock.)

"hang" a thousand or more feet above the bottoms of the main valleys, not uncommonly producing beautiful waterfalls.

Scouring the bottoms of valleys characteristically produces basins which, when filled with water, make the picturesque lakes for which certain mountain regions are famous (Fig. 189). If a vigorous valley glacier extends into the sea it may erode the bottom deep below sea-level. After the glacier has melted, such a long, deep, steep-sided embayment is called a *fiord* (Fig. 190).

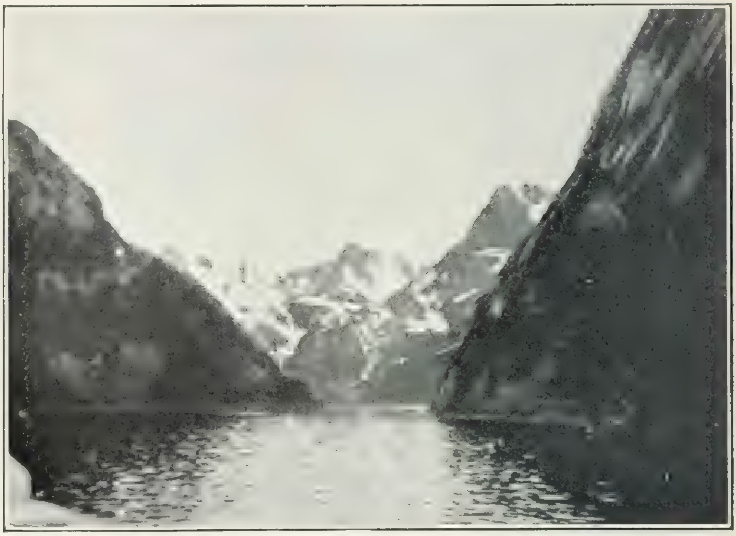


Fig. 190. — Troldfjord, Lofoten Islands, coast of Norway. (R. T. Chamberlin.)

Ice-caps which overspread the surface irrespective of valleys and hills tend to reduce the angularities of the surface. Hills and ridges are cut down and smoothed (Figs. 191 and 192), rendering them in general more gentle on the iceward slopes (*stoss sides*), while the lee slopes retain more of their original steepness. And further the lee side may be even steepened by plucking. But on the other hand valleys parallel with the direction of ice movement are deepened, as shown by the basins of the Great Lakes and the Finger Lakes of New York. Rounded bosses of rock (*roches moutonnées*) result from strong erosion of originally rough rock surfaces (Fig. 193).

Positions in which Debris is Carried. Debris is seen to be carried by a glacier (1) at, or near, its bottom, *basal* or *subglacial;* (2) up in the body of the ice, *englacial*, and (3) on the surface, *superglacial.* It is evident from the foregoing discussion that subglacial material has been picked up from the bed or has found its way to the bottom through crevasses, etc., while the superglacial load has fallen upon the ice from adjoining cliffs or steep slopes or has been blown there as dust by the wind. The other, or englacial, load requires further consideration. If the glacier passes over a prominence, material may be plucked or rasped from its upper part and carried along at that level (Fig. 194). It may also reach an englacial position by falling from the surface into crevasses or holes.

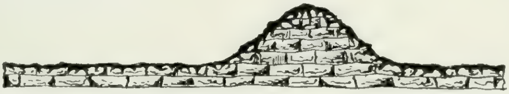


Fig. 191. — Diagram representing a hill unworn by ice, and the irregular contact of soil and rock.

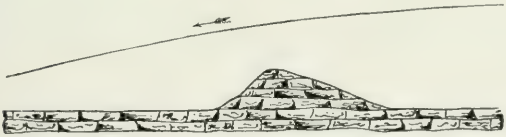


Fig. 192. — Diagram showing the effect of glacial wear on a hill such as is shown in Fig. 191.

Furthermore load is transferred from a basal to an englacial and then to a superglacial position in many cases by upward movement. Such transfer is the more remarkable because the specific gravity of rock is two and a half or three times that of ice, so that the normal tendency of rock is to sink in ice. The upward movement of debris is the result of the shearing which is prevalent in large active glaciers. The upper portion of a glacier, encountering less resistance than the lower portion, shears forward over the retarded lower portion. This takes place where the ice in front of the plane of shearing lies at a lower level than that behind, as where the surface of an upland falls off into a valley, or where a boss of rock shelters the ice in its lee from the thrust of the over-riding ice (Fig. 195).

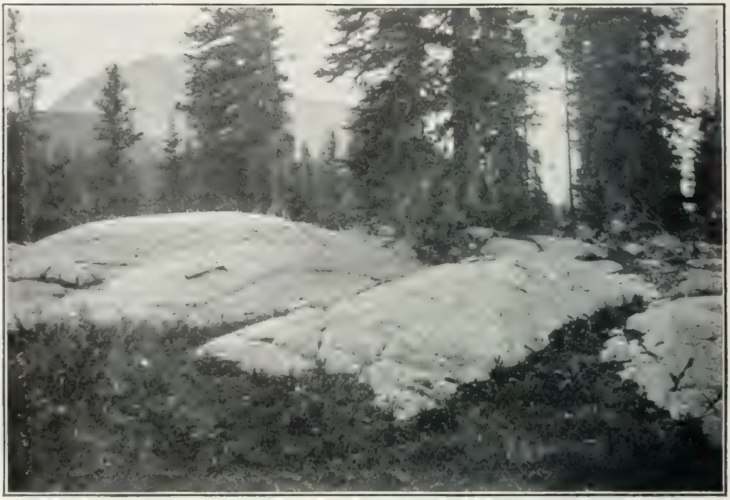


Fig. 193. — Roches moutonnées on Engineer Mountain, Colo. (Hole, U. S. Geol. Surv.)

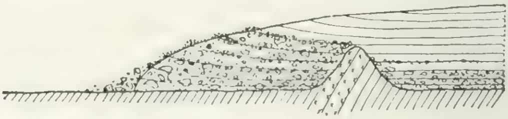


Fig. 194. — Diagram illustrating one way in which a glacier gets englacial material.

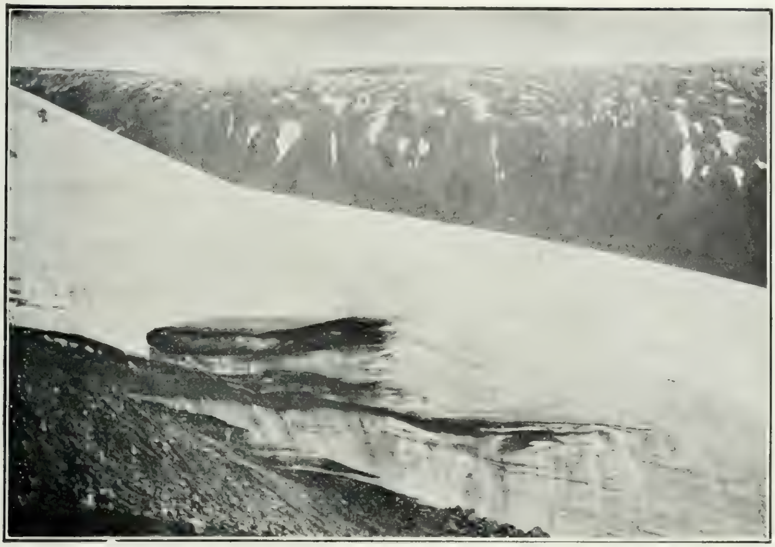


Fig. 195. — Taking debris from a protuberance of the bed.

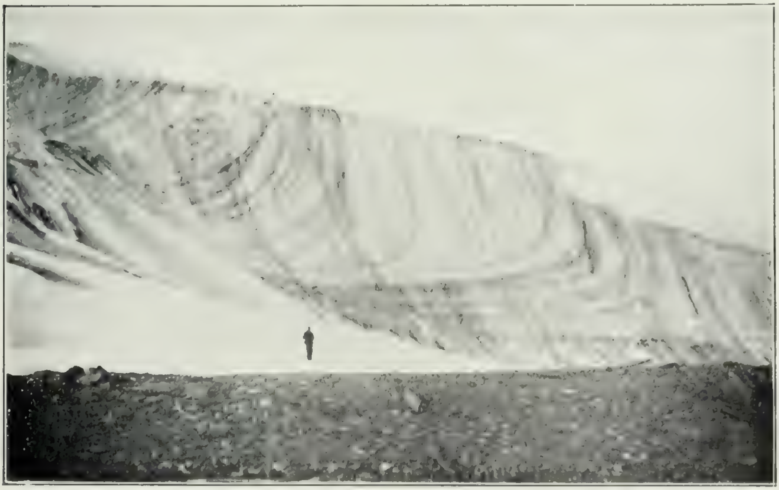


Fig. 196. — End of a North Greenland glacier, showing the upturning of the layers of ice at the end. This structure is common in North Greenland. At one point, a few stones are seen on the surface of the ice where an upturned layer comes to the surface.

Shearing takes place also along the borders of practically all valley glaciers, where the margins are less free to move and the ice behind, because of strong thrusting, slides up and over. Along the borders of glaciers the layers are normally turned up, as shown in Figure 196. In this figure, however, the layers are even overturned owing not only to the resistance of frozen morainic or other material beneath and in front of the edge, but probably also to the effect of the great rigidity of the outer part of the ice, due to the low external temperature during the larger part of the year, while the interior, with its higher temperature, moves more rapidly. Where the shearing planes turn up at the end or side of a glacier, basal and englacial debris may be carried to the surface by actual upward movement, and a terminal or lateral moraine, or a series of such moraines, is sometimes aggregated *where the upturned layers of ice outcrop at the surface* (Fig. 197). That much of the material of these moraines was originally basal is abundantly demonstrated in many cases by the bruised and scratched condition of the boulders and pebbles, and by the nature of the material itself. A valley glacier which has resulted from the junction of two tributary glaciers commonly exhibits upturning of the ice along a longitudinal zone well back from the lateral margins, and the material so borne to the surface in such a zone gives rise to a medial moraine. This medial upturning of the ice is a

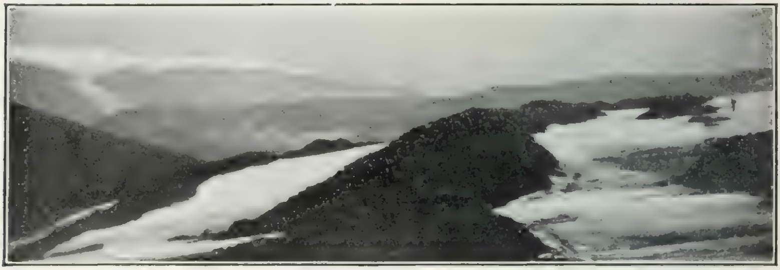


Fig. 197. — Surface terminal moraines due to upturning. Edge of the ice sheet, North Greenland.

persistent feature because the two parts of the composite glacier maintain their integrity for a long distance beyond the place of junction.

DEPOSITION OF DRIFT

All the material which has been handled by a glacier is known as *glacial drift*. Glacial drift is of two more or less distinct varieties: (1) Where debris-laden ice melts, much of the load is deposited directly. It retains the heterogeneity in size, shape, and kind of material, that characterized the debris while in the ice. Such drift, made up predominantly of "rock-flour" in which sand grains, pebbles, cobbles, and boulders are scattered irregularly, is un-

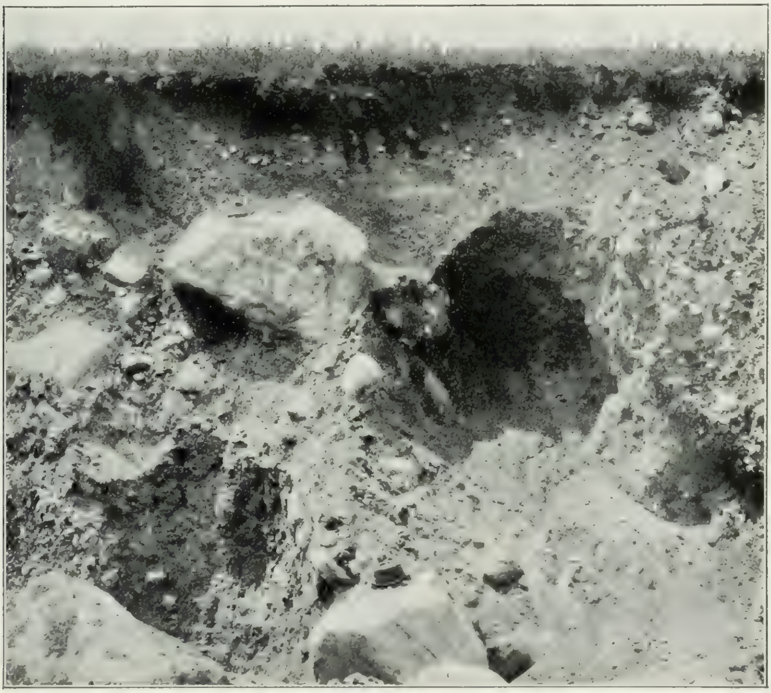


Fig. 198. — Section of till showing its heterogeneity.

stratified and is known as *till* (Fig. 198). (2) In addition, the water resulting from a melting glacier carries, sorts, stratifies, and deposits other portions of the debris to form *glacio-fluvial drift* (Fig. 200).

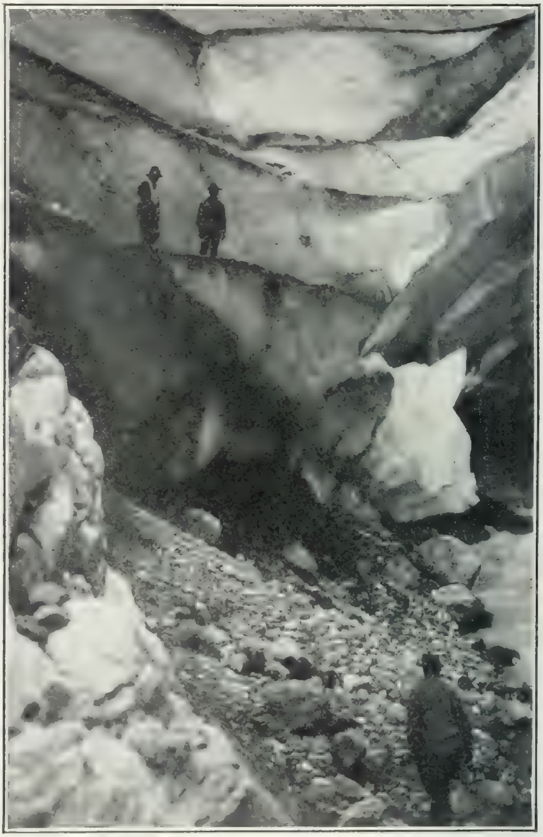


Fig. 199. — Debris being deposited at foot of glacier. British Columbia. (Porter.)

UNSTRATIFIED (TILL) DEPOSITION

Till is deposited in two more or less distinct situations: (1) under the body of the glacier, and (2) near, under, or at the edge of the glacier.

(1) Even beneath moving ice a certain amount of deposition is constantly taking place. Rapid erosion at one place may result in such loading of the ice that changed conditions, such as lower grade, irregularities of bed, more rapid melting, etc., farther on, cause deposition of some of the material. The topographical result in the case of a continental glacier is a gently rolling plain, *ground moraine*, characterized by long, gentle hills (swells), irregularly distributed, with broad gentle depressions (swales) between. The topography may be characterized as *swell-and-swale*. The thickness of ground moraine varies from 500 or 600 feet to little or nothing, for there are numerous uncovered spots of bed-rock even in a glaciated area. In places ground-moraine shows hills of peculiar oval shape, elongated in a direction parallel with the ice motion. Such hills of unstratified drift are called *drumlins*. They are steeper on the iceward side than on the lee, and their shape suggests a depositional rather than an erosional origin, steeper because of deposition *against* the

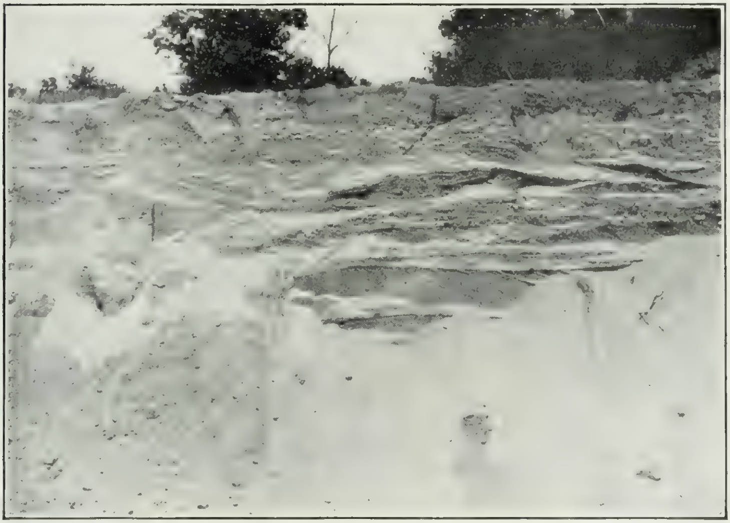


Fig. 200. — Glacio-fluvial drift. (Alden. U. S. Geol. Surv.)

"upstream" end. The fact that they are prevalent chiefly where glaciers have spread and crevassed longitudinally, suggests that drumlins may owe their origin to deposition of drift either beneath longitudinal crevasses or under special conditions of glacier-spreading.

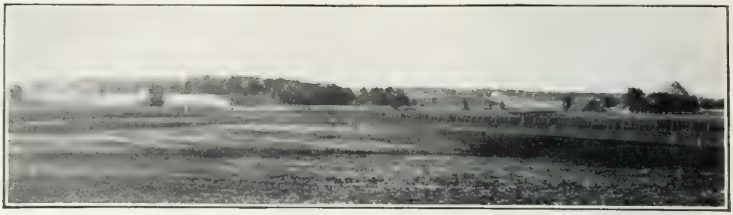


Fig. 201. — Gently rolling ground moraine topography. Barrington, Ill. (MacClintock.)

(2) At the edge of an ice-cap, or at the end of a valley glacier, where the ice completely melts, all the load is liberated and much of it accumulates as a thickened belt, or ridge of drift, known as a *terminal moraine*. The characteristics of a terminal moraine become significant when we consider in some detail the conditions of its deposition. In the first place, the ice-edge remains stationary only when the forward motion of the ice is exactly balanced by melting. But the two are equal only at times. Melting is more rapid

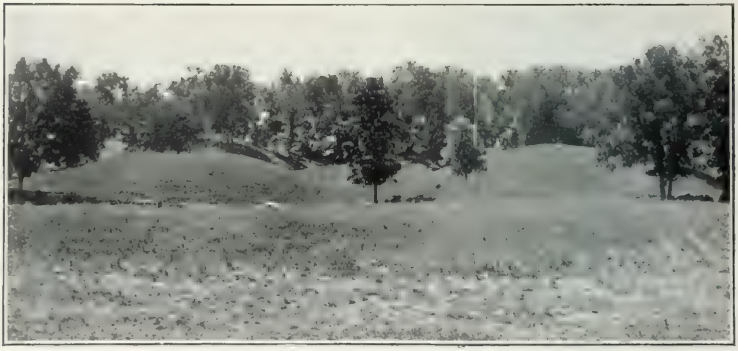


Fig. 202. — Terminal moraine topography. (Alden U. S. Geol. Surv.)

in summer than in winter, in some summers than in others, and furthermore, the ice at some places and at some times is less dense than at others, and so melts faster. In the second place, the speed of ice movement varies from time to time and from place to place. In consequence the edge of the ice fluctuates a great deal. In addition, large cracks, re-entrants and embayments extending back into the ice, as well as large masses of ice isolated by melting and lying beyond the actual edge, serve to render the glacier margin exceedingly irregular. In consequence of this and the masking of the ice under thick debris, it is impossible in many places to determine the actual margin of the glacier. And in the third place, the ice contains differing amounts of debris from place to place and from time to time. When a certain amount of heavily laden ice melts, much drift is deposited, while an adjacent area of clean ice may liberate only a small amount of debris. Or, much may be deposited at one time and little at another time in the same place.

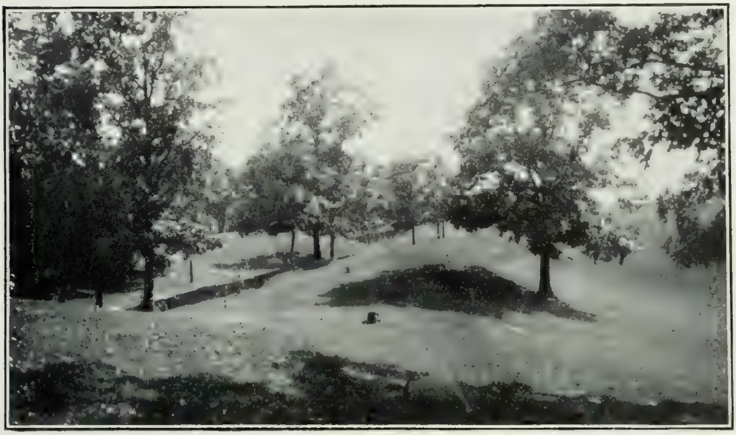


Fig. 203. — Terminal moraine topography. (MacClintock, Ill. State Geol. Surv.)

Such irregular deposition of debris at and near the edge of the ice, results in a thick accumulation of drift made up of many dump-piles. In many places the surface consists of a maze of small, hummocky knob-like hills and irregularly distributed kettle-like depressions giving rise to the term *knob-and-kettle* topography (Figs. 202 and 203). Many of these depressions contain small lakes. In other parts of the same moraine and in other moraines, however, the topography is that of a ridge or a raised belt of thick drift without the characteristic knobs-and-kettles.

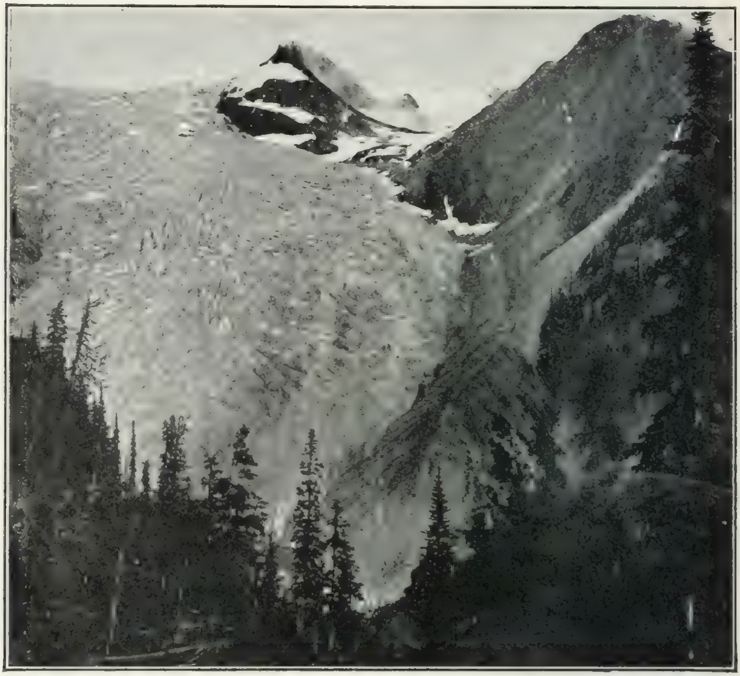
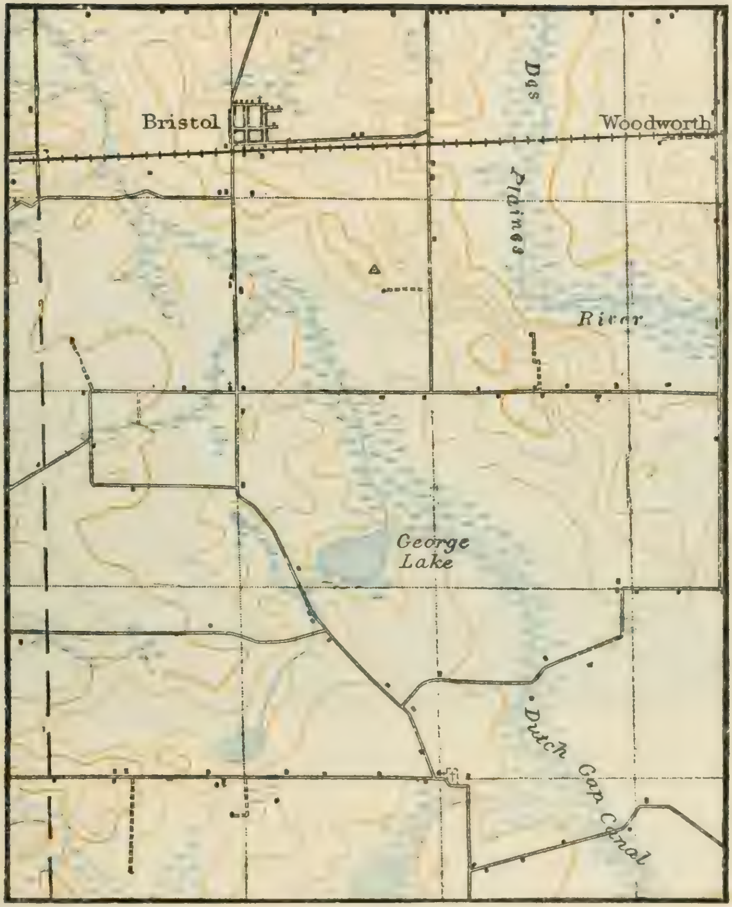


Fig. 204. — Illecillewaet Glacier, British Columbia, showing lateral moraine.

In valley glaciers ice moves not only down the valley but from middle to sides. This becomes self-evident when it is recognized that melting along the sides would narrow the glacier to a very small tongue of ice if it were not for this spreading motion. The result of this lateral motion is to bring debris to the sides of a glacier where it accumulates

PLATE XVII


Bristol
Woodworth
Des Plaines River
George Lake
Dutch Gap Canal

Characteristic surface of a glaciated plain, showing marshes, ponds, and lakes. Southern Wisconsin. Scale, about 1 mile per inch. (Silver Lake, Wis., Sheet, U. S. Geol. Surv.)

to form lateral moraines (Figs. 204 and 205). Lateral moraines consist of material in part dumped over the edge of the glacier from the superglacial lateral moraine, in part englacial debris liberated by melting, and in part drift pushed or dragged along by the basal portion of the ice. Some of the lateral moraines of ancient valley glaciers, such as those of the Uinta, Wasatch, and Bighorn mountains are

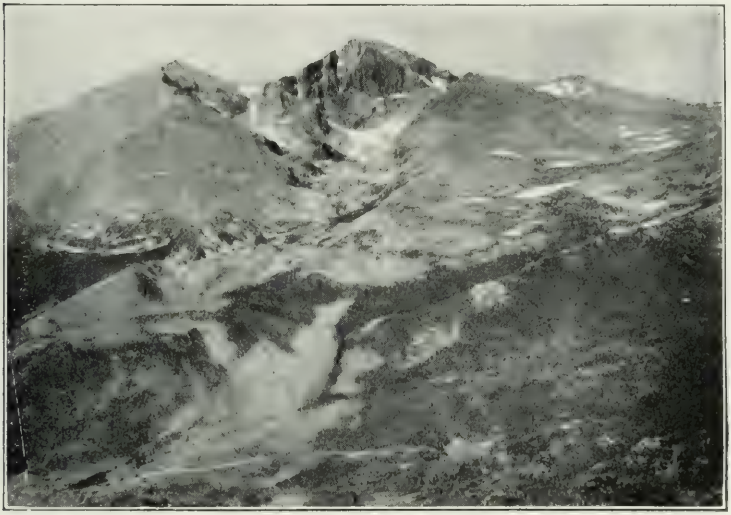


Fig 205. — Long's Peak, Colo., showing cirque and moraines. (W. T. Lee, U. S. Geol. Surv.)

several hundred and even as much as a thousand feet high. There is in northern Italy a lateral moraine measuring 2000 feet in height.

STRATIFIED (GLACIO-FLUVIAL) DEPOSITION

The continuous but unequal waste of glaciers by melting has already been referred to. The streams to which the melting ice gives rise are heavily laden with glacial debris, gravel, sand and silt and hence are, as a rule, aggrading streams. Both the coarse and fine material is carried for a greater or less distance, but soon the gravel-stones are dropped, later the sand, and eventually the silt and finely comminuted material. Thus it happens that glacio-fluvial drift is always assorted; and most completely assorted farthest from the glacier. Furthermore, unequal melting gives rise to unequal volume of glacial waters and hence unequal carrying capacity from time to time. Stratification of the deposits results, layers of coarser material being deposited during the time of much and strong water, and layers of fine material during low water.

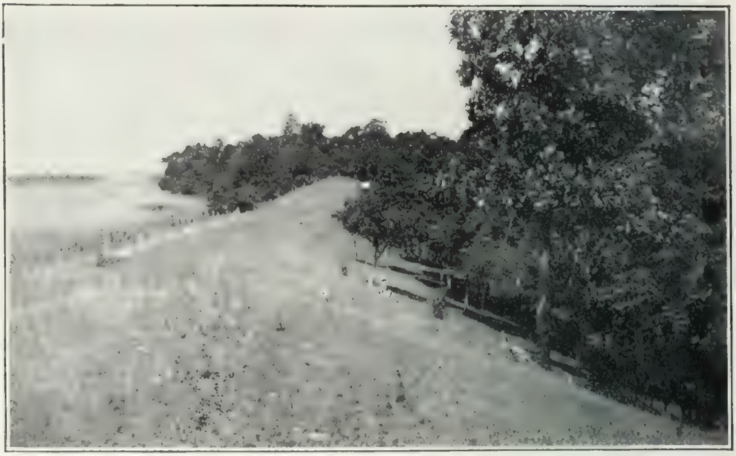


Fig. 206. — Bridgewater Esker, Rice County, Minn. (R. T. Chamberlin.)

During the advance of the glacier, its maximum extension, and its retreat, glacial waters not only deposit stratified material, but rework and modify part of the drift left directly by the ice. Till and assorted drift (largely gravel and sand) are consequently closely associated in much of the area covered by a glacier.

Deposition by glacial waters takes place in three different situations which overlap to some extent as the margin of the glacier advances or retreats. These situations are (1) back under the ice, (2) at or near its edge, and (3) beyond its edge.

(1) Where streams of considerable size form tunnels under the ice, the tunnels may become more or less filled with water-worn debris, and when the glacier melts, the aggraded channels appear as ridges of gravel and sand known as *eskers* (Fig. 206). Some eskers are so strikingly regular as to resemble railway embankments in general appearance. They may be many miles in length and several scores of feet in height.

(2) At the mouths of ice tunnels or ice channels, especially where they end against terminal moraines, and in re-entrant angles of the edge of the ice, glacio-fluvial material is liable to be assembled in quantity, giving rise, after the adjacent ice has melted, to steep hills and short ridges known as *kames* (Fig. 207). Bearing in mind the complexity of the ice margin, it may be safely said the kames are formed in, among, and under the edge of the ice.

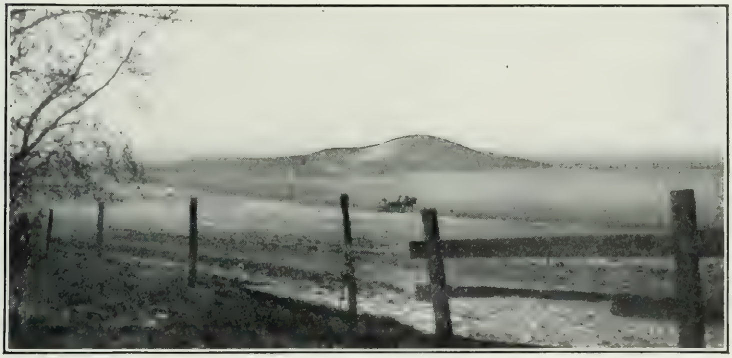


Fig. 207. — An isolated kame 3 miles northeast of Elgin, Ill. (MacClintock, Ill. State Geol. Surv.)

(3) In the case of a valley glacier, the heavily laden outwash waters descending the valley aggrade its bottom to make a type of alluvial plain, known as a *valley train* (Fig. 208). In the case of an ice-cap, if the waters find a valley they likewise aggrade its bottom to form a valley train. But if they fail to find a valley each stream builds

an alluvial fan. By the coalescing of many such fans, an *outwash plain*, sloping gently away from the terminal moraine, is developed (Fig. 209). The material is naturally coarser near the moraine, becoming finer farther and farther away. Where glacial drainage issues into standing or slow-flowing water, deltas are commonly built.

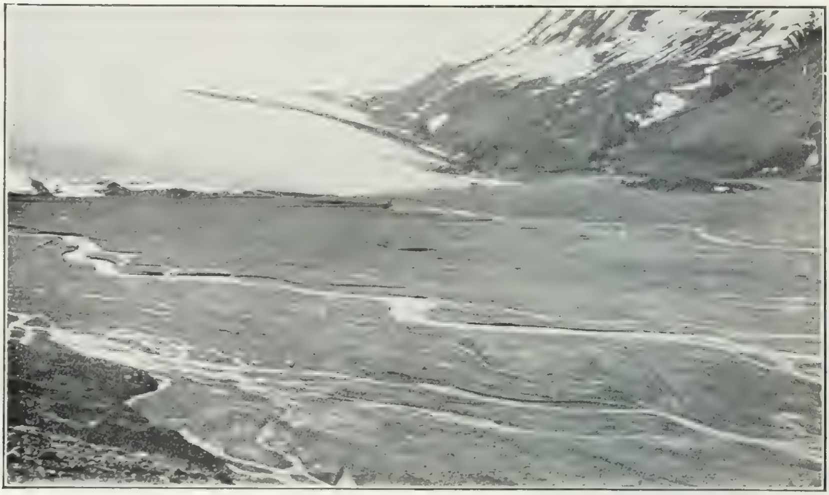


Fig. 208. — Streams from Hidden Glacier, Alaska, making valley train.

This outwash material may bury outlying masses of ice left during recession of the ice-front from an earlier more advanced position. Upon subsequent melting of the buried ice-blocks pits are formed below the surface of the outwash plain (*pitted plain*). Similarly blocks of ice buried under the accumulations of the terminal moraine upon melting give rise to a "let-down" topography in which gigantic pits and depressed areas are characteristic.

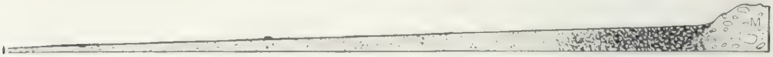


Fig. 209. — Diagram to illustrate the profile of an outwash plain, and its relation to terminal moraine. M is the moraine.

As a result of all these phases of water-work associated with ice-work at all stages of glacial growth and decay, a considerable part of the drift is assorted and stratified. Stratified drift may lie above or below unstratified drift, or the two may be interbedded in almost any conceivable way.

GLACIO-NATANT DEPOSITION

Where glaciers advance into water, the depth of which approaches their thickness, blocks of ice are broken off and float away as *icebergs*. Some of them are 100 to 200 feet high above the water, which means that they may extend many times as far below the surface of the water. As the bergs set sail they carry with them the debris they contained while still part of the glacier. This debris may be carried for several hundred miles before melting liberates it. Material carried in this way is, however, not of great quantitative importance in any given area of deposition.

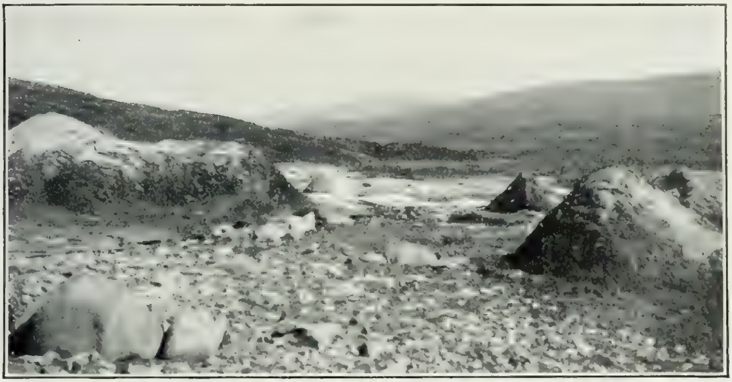


Fig. 210. — Masses of ice partly covered with outwash debris. (P. S. Smith, U. S. Geol. Surv.)

RÉSUMÉ

1. Valley glaciers leave the following characteristic marks: (*a*) U-shaped valleys above grade (modified youth-

ful valleys); (*b*) hanging tributary valleys; (*c*) cirques at the heads of the valleys; (*d*) striated, smoothed, and polished bed-rock; (*e*) lakes; (*f*) moraines, lateral and terminal, made of glacial drift; (*g*) valley trains; and (*h*) perched boulders.

2. Ice-sheets also leave characteristic features: (*a*) striated, grooved, and smoothed bed-rock; (*b*) drift of two kinds: unstratified *till* and stratified glacio-fluvial material; and (*c*) distinctive topography of ground moraine, terminal moraine and outwash, with the lakes, ponds, marshes, and haphazard drainage typical of glaciated country.

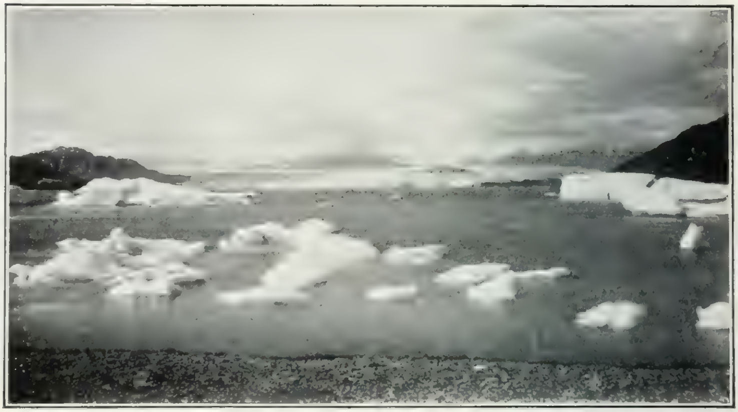


Fig. 211. — Icebergs in a Greenland fiord.

ANCIENT ICE-SHEETS

At several times in geologic history glacial climates have given rise to extensive ice-sheets on the various continents. The most recent of these periods of widespread glaciation is know as the *Glacial Period*. During this period more than 4,000,000 square miles of North America and a large part of northern Europe were covered by glacial ice.

Not only were existing glaciers in the mountains much larger than at present, but many lay in mountain valleys which are today too warm to accumulate perennial snows. The North American ice-sheet (Fig. 212) apparently developed from three centers (several small ones are not considered here); one east of Hudson Bay, one west of it, and one in the western mountains. The spreading of ice-sheets from these different centers may not have been precisely contemporaneous. But, setting details aside for the moment, ice from the first two centers spread far south into

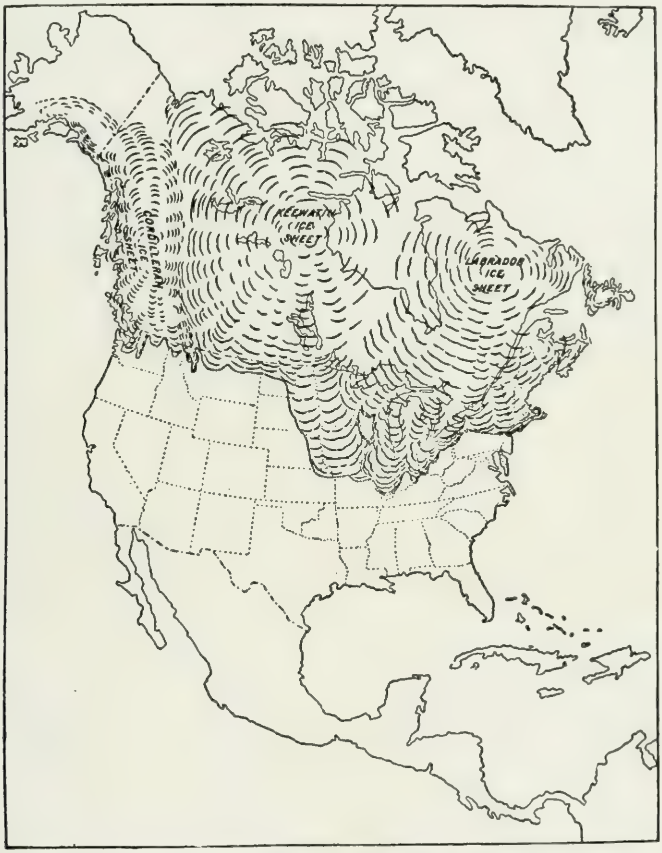

CORDILLERAN ICE SHEET
KEEWATIN ICE SHEET
LABRADOR ICE SHEET

Fig. 212. — Sketch-map showing the North American ice sheets with their centers of spreading.

the United States. The line of maximum extent may be drawn from Long Island and New York City on the east, across Pennsylvania, down the Ohio River Valley, across the southern tip of Illinois, up the Missouri, and westward to Puget Sound. Within this glaciated region, an area of about 10,000 square miles, mainly in southwest Wisconsin,

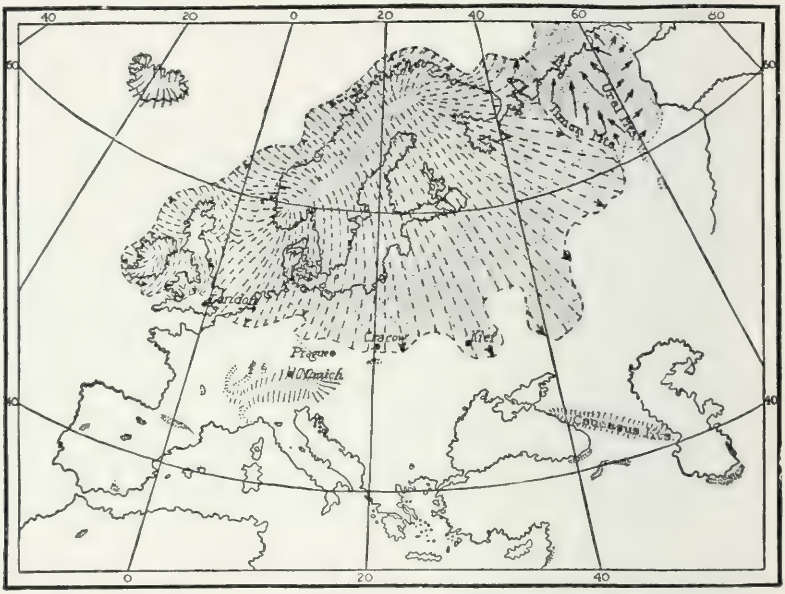


Fig. 213. Sketch-map showing area of Europe covered by the continental glacier at the time of its maximum development. (Jas. Geikie.)

was never covered by the glaciers. Because of the absence of drift, it is known as the *Driftless Area.*

In Europe, likewise, existing glaciers extended much farther than at present, other mountain glaciers were present where today none are to be found, and a great ice-cap covered the northwestern part of the continent (Fig. 213). While the main center of this ice-cap was in the Scandinavian highlands, there were local centers in Scotland and the Urals from which the ice spread.

The history of these great ice-sheets was not a simple one, but was made up of several successive ice advances and retreats. In North America the first advance was followed by retreat of the glaciers to very small size or possible total disappearance. The drift was exposed for so long that deep soil developed upon its surface, and plants and animals took possession and flourished in a climate even warmer than that of today. Glacial conditions then returned, for a second ice-sheet formed and advanced over approximately the area covered by the first. During this second advance the soil and vegetation of the first drift sheet were buried in places and preserved as *interglacial deposits.* This and other criteria have been used to differentiate a third, fourth, and possibly a fifth epoch of ice-advance within the Glacial Period. The Glacial Period in Europe has been found to have a somewhat similar history.

CHAPTER VI

THE OCEAN

GENERAL FACTS

As pointed out in Chapter I, the oceans cover about three-fourths of the surface of the lithosphere, while the abysmal portions occupy about two-thirds. This great body of water lies for the most part in the depressed areas called *ocean basins.* The term basin, however, suggests a concave depression, but the true shape of the bed of the oceans is very different from this familiar image. Figure 214 shows that the ocean basins are in general convex upward, being concave only locally near the margins or in certain very deep portions. The water of the oceans might be likened to the film which would adhere to the surface of a ball after it had been dipped into water.

The surface of the sea is known as sea-level. It is, in general, the curved surface of an oblate spheroid, but local irregularities distort it somewhat. High land masses or portions of the lithosphere denser than the average so attract the water that the surface is raised in their vicinity. It follows that the higher and more massive the adjacent land, and the denser the rock, the stronger will be the attraction, hence the higher the sea-level.

The oceans are all connected at the surface, making them in a sense one, though they are known by different names. The Southern Ocean encircles the globe in about 60° South Latitude. From it three more or less irregular arms the Indian, the Atlantic, and the Pacific oceans, extend northward to connect, in the case of the last two, with the Arctic Ocean.

The depth of the sea averages close to $2\frac{1}{4}$ miles for the Atlantic, $2\frac{3}{4}$ miles for the Pacific, and $2\frac{1}{4}$ miles for the Indian Ocean. There are tracts, notably deeper than these average figures, known as *deeps*. The deepest point so far reported lies about 50 miles off the east coast of Japan in the Tuscarora Deep. The Japanese man-of-war, the *Manchu*, let out its sounding wire 32,644 feet long, without striking bottom. The deepest place in the ocean previously reported was 32,113 feet off the coast of Mindanao, Philippine Islands. The Pacific Ocean contains other deeps of note; 31,614 in the Ladrone Deep, 30,930 in the Aldrich Deep northeast of New Zealand, and 25,000 feet off the coast of Chili. In the Atlantic Ocean a depth of 27,922 feet is known just east of Haiti, and 27,366 feet in the Blake Deep north of Porto Rico. The deeps of the oceans all have pronounced elongation parallel with the nearest coast or submarine ridge.

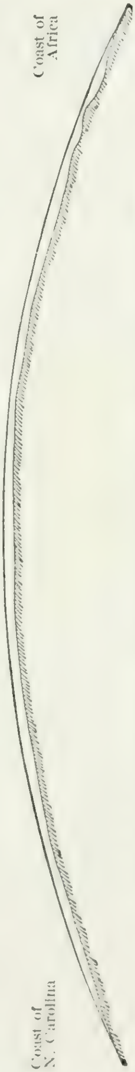


Fig. 214. — Diagram illustrating the true character of the ocean basin.

Volume and Mass. It has been computed, from the depths and area, that the ocean contains 323,722,150 cubic miles of water, or about 15 times the volume of the land above sea-level. If the land above sea-level were eroded and the material deposited in the ocean basins it would raise the sea-level about 650 feet; or if the continental masses were cut down and the ocean basins filled up bringing the surface of the lithosphere to a common level, a universal ocean about 9000 feet deep would cover the entire lithosphere.

The mass of the hydrosphere is 265 times that of the atmosphere, but only about 1/4500 part of that of the lithosphere

Topography of the Ocean Bottom. If the water were removed from the bed of the ocean so that it could be seen as the land is, its most impressive feature would be its *monotony.* The hills and valleys which, in all their multitudinous forms, give the land its most familiar features, are essentially absent. A large part of the ocean bottom is so nearly flat that the eye would not detect its departure from planeness.

The reason for this profound difference between the land and sea bottom is readily found. The dominant processes which shape the details of the surface of the land are degradational, and though the final result of degradation is flatness (base-level), the immediate result is roughness. In the sea the dominant processes are aggradational, and tend to monotonous flatness.

The flatness is, however, broken by rises to (1) volcanic peaks, (2) ridges of mountainous heights, (3) extensive swells or tracts covered with water of moderate or shallow depth, and by declines into the deeps. Of these slopes one as steep as one mile in twenty is very rare, so that a railway train could be run directly up practically all the "steep" slopes on the ocean bed, so far as these are now known, except perhaps some of volcanic origin and some produced by local fracturing.

Composition of Sea Water. Every 1000 parts of sea-water contain from 31 to 37 parts by weight of mineral matter in solution. Three-fourths of the mineral matter is common salt (sodium chloride), but other salts are also important. The percentages of the principal solids are shown in the following table.[1]

Sodium chloride...............	77.758
Magnesium chloride...........	10.878
Magnesium sulphate..........	4.737
Calcium sulphate.............	3.600
Potassium sulphate...........	2.465
Calcium carbonate............	0.345
Magnesium bromide..........	0.217

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[1] Dittmar, Challenger Reports, Physics and Chemistry, Vol. I, p. 204.

Other mineral substances are found in small quantities. Gases such as oxygen, nitrogen, and carbon dioxide are present in quantity. It is estimated that there is 18 to 25 or more times as much carbon dioxide in the sea as in the atmosphere. The presence of mineral matter in solution makes sea-water slightly heavier than fresh water, its specific gravity being about 1.026.

The volume and composition of sea-water being known, the amount of mineral matter which it contains may readily be calculated. Assuming the average specific gravity of the mineral matter to be 2.5, the 3.5% (nearly) by weight becomes about 1.4% by volume, and 1.4% of 323,722,150 is 4,532,110 cubic miles. This represents the aggregate volume of mineral matter in the sea if it were precipitated and compacted so as to have an average specific gravity of 2.5. This amount of mineral matter would cover the ocean bottom to a depth of about 175 feet. Its volume is equal to about 20% of that of all the lands now above sea-level.

A large part of the mineral matter of the sea has come from the land, having been extracted from it by ground-water, and carried by rivers to the sea. But the mineral matter of the sea gives no more than a hint of the importance of the solvent work of water in the general processes of rock decay, for most of the mineral matter carried from the land to the sea in solution is extracted from the sea-water about as rapidly as it is supplied. Calcium carbonate, for example, is about 20 times as abundant as sodium chloride in river-water, but it is only 1/225 as abundant in sea-water, due to the fact that the calcium carbonate is used by animals and plants to make shells, skeletons, etc., while the salt remains in solution.

MOVEMENTS OF SEA-WATER

The movements of sea-water fall into several categories. There is (1) a general circulation of sea-water, determined chiefly by (*a*) differences in density in the sea-water, (*b*) differences of level, and (*c*) movements of the atmosphere; (2) periodic tidal movements; and (3) nonperiodic movements due to occasional causes, such as earthquakes, volcanic explosions, landslides, etc.

For present purposes, all movements of the sea-water may be grouped into two main classes — (1) waves, with the undertow and the littoral currents they generate, and (2) ocean-currents.

Waves. The most common waves are those generated by winds. During the passage of a wave, each water particle affected by it rises and falls, and moves forward and backward, describing an orbit in a vertical plane. Such waves are known as waves of oscillation. If the passing wave is a swell, the orbit of the particle is closed and is either a circle or an ellipse; but in the case of a wind-wave the orbit is not closed, for in such a wave the water, as well as the undulation, moves forward. On the crest of the wind-wave each particle of water moves forward, and in the trough it moves less rapidly backward, and the excess of the forward movement over the backward gives the water a slight residual advance. This advance is the initiatory element of current. As a result of this advance, the upper layer of water is carried forward, with reference to the layer below, in the direction toward which the wind blows. The waves of any considerable or long-continued wind, therefore, generate a surface current in the same direction as the wind.

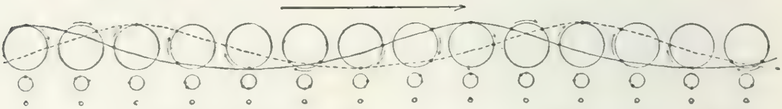


Fig. 215. — Wave motion produced by the circular revolution of the water particles.

Wave motion is propagated indefinitely downward, but the amount of motion diminishes rapidly with increasing depth (Fig. 215). Engineering operations have shown that submarine structures are little disturbed at depths of 15 feet in the Mediterranean, and 25 feet in the Atlantic. On the other hand, rock material as coarse as gravel, which is transported by rolling on the bottom, is not infrequently carried out to depths of 50 feet, and fine sediment, like silt, is disturbed at still greater depths.

When a wave of oscillation approaches a shelving shore, its habit is changed. The forward velocity of the undulation (wave-form) is diminished due to shortening of the wave-length (measured from crest to crest) because of diminishing depth of water. This greater retardation of the shoreward waves than those behind allows the latter to close up on those ahead, causing a bunching. At the same time each wave gets higher and higher, the orbits of the individual particles becoming larger and larger owing to the concentration of energy in the greater displacement of the fewer particles. Since, owing to the bunching, the frequency, i.e., the number of waves passing a given point in a given time, remains constant, meaning that the water particles must describe their larger orbits in the same time as they did their smaller ones and hence have acquired greater orbital velocity. This orbital velocity of the parti-

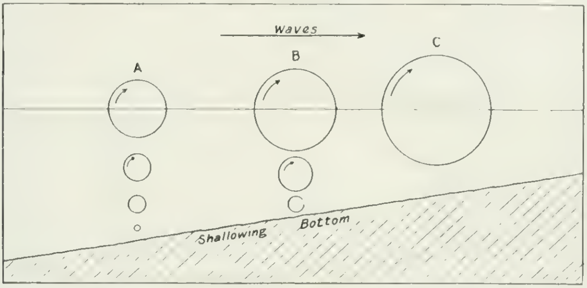


Fig. 216. — The energy of wave-motion at A is imparted to fewer particles at B resulting in their greater displacement. At C the energy has been concentrated into still fewer particles producing still greater displacement (higher waves).

cles becomes greater and greater while the forward motion of each wave becomes slower and slower. When the orbital velocity of the particles in the upper part of the water finally becomes greater than the forward propagation of the wave, the crest curves forward and breaks.

Waves of a given height in a given wind break in about the same depth of water, and the line along which incoming waves break is the line of *breakers*. The line of breakers is in deepest water and farthest from shore when the waves are strong. When the waves of oscillation break, they give rise to smaller waves of a different character in which the water moves bodily forward and hurls itself upon the beach (*waves of translation*). The return of the water thrown forward in the crests of the waves is accomplished by a current along the bottom called *undertow*. The undertow is sensibly normal to the coast when uninfluenced by oblique waves, and is efficient in removing the products of erosion.

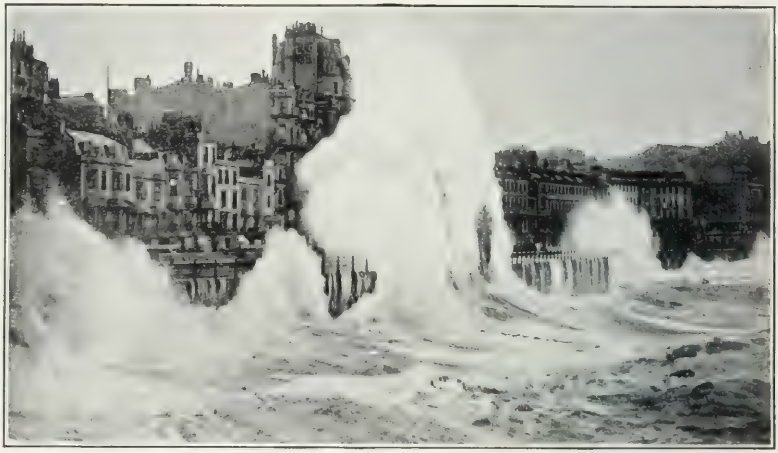


Fig. 217. — Shore wave dashing on east wall of Hastings. (From Wheeler's *The Sea Coast;* by permission of Longmans, Green and Company.)

When waves advance on the shore obliquely, a shore-current is developed as illustrated by Figure 218, where *ab* represents the direction of the incoming wave, *bc* the direction of the shore (or littoral) current, and *bd* the direction of the undertow. Where they strike the borders of land, the wind-waves therefore generate two other movements, the undertow and the littoral current. Any particle of water near shore may be affected by any two or by all three of these movements at the same moment. The effect of the littoral current and undertow is to give a particle of water on which both are working a direction between the two, as *be*. The effect of the other combinations can be readily inferred. These various combinations are of consequence in the transportation of debris.

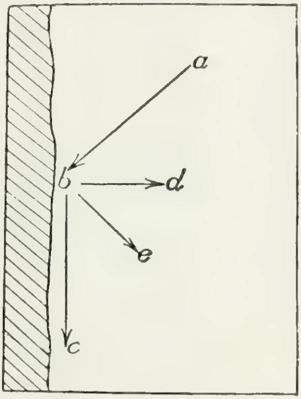


Fig. 218. — Diagram showing relative directions of wave, undertow and shore-current.

WORK OF THE WAVES

Erosion. It has been estimated that the force of the average winter waves on the coasts of the British Isles is about a ton per square foot, while storm waves on exposed places reach a force of as much as three tons per square foot. The power of such waves pounding against an exposed coast is stupendous. In the dash of waves against the shore, the wear is effected chiefly by the impact of the water and of the debris which the water carries. Lesser results are accomplished in other ways.

Where the land at the margin of the water consists of unconsolidated material, or of fragmental material but slightly cemented, the dash of the water is sufficient to displace or erode it. If weak rock is associated with resistant rock within the zone of wave-work, the removal of the former may lead to the disruption and fall of the latter, especially when the weak rock is washed out from beneath the strong. The impact of the water is competent also to break up and remove rock which was once resistant, but which has been weakened by weathering. Rock affected by joints is likewise attacked with success, for the blocks bounded by joints may be loosened and literally quarried out. On the other hand, waves of clear water, even when their force is several tons to the square foot, have little effect on rock which is thoroughly solid.

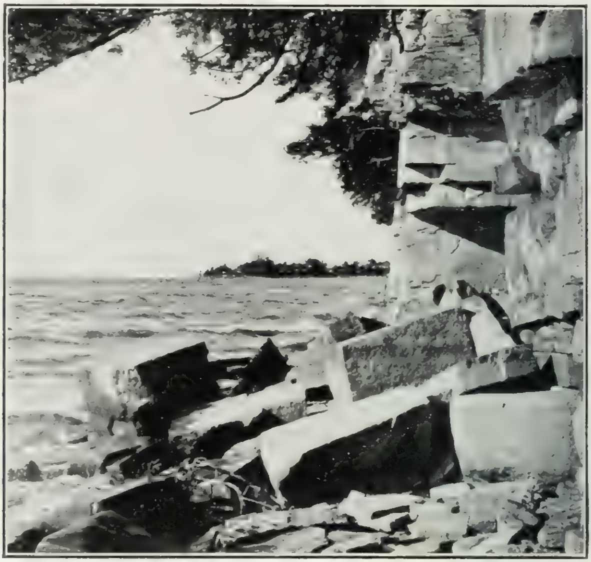


Fig. 219. — Angular blocks of rock which have fallen from the cliff above, as a result of undercutting by the waves; Grand Island, Lake Champlain. The rock is Black River limestone. Although from the shore of a lake instead of the sea, the principles illustrated are the same. (Perry.)

The impact of the waves is generally reinforced by the detritus they carry. Sand, pebbles, and such stones as the waves can move, are used as weapons of attack, both against the shore and against one another. Masses of rock too large for the waves to move (Fig. 219) are worn by detritus driven back and forth over and against them, and in time are reduced to movable dimensions. They then become the tools of the waves and, in use, are reduced to smaller and smaller size. The products of this wave-wear are boulders, cobbles, pebbles, sand, silt and mud. The silt and mud, held in suspension in agitated water, are carried out beyond the range of the breakers and settle in water so deep as not to be effectively agitated at the bottom. The sand, because it settles faster, is not carried out so far before it is deposited. Thus one generation of boulders after another is worn out and the comminuted products are carried out from the immediate shore and deposited in deeper water.

The effectiveness of waves depends (1) on their strength, (2) on the concentration of their blows, (3) on the nature of their tools and (4) on the resistance of the shore-rock. 1. The strength of waves is dependent on the strength of the winds (or other generating causes), the depth of the water and the distance they have traveled (*fetch*). For a given coast the effectiveness of wave-erosion is determined less by the average strength of waves than by the strength of the storm-waves. Exceptional storm-waves are known to have moved blocks of rock exceeding 100 tons in weight. 2. Waves are most effective when concentrated, as on bold coasts bordered by broad expanses of deep water, for here the force of the waves is almost wholly expended near the water-line; where shallow water borders the land, the force of the waves is expended over a greater area. 3. And as we have seen in the case of running water, erosion is most rapid where the tools are of the right size, shape and kind. 4. The effect of rock resistance is obvious.

The less familiar phases of wave-work are accomplished by hydraulic pressure, compressed air, shore-ice, etc. When the water of a wave is driven into an open joint or a cave, the hydraulic pressure may be so great as to break the rock if it is weak. When water is driven with force into a cave the compression of the air may be great if the wave is high enough to close the entrance. Some of the air may be expelled through cracks, so that when the water runs out of the cave, the air within may be greatly rarefied, while that above exerts its normal pressure. In either case the roof of a cave, if it is weak, may be broken. At certain seasons of the year, especially during the spring, waves make destructive use of the ice which is then breaking up, but it is only in high latitudes that sea-ice is of importance in this way. In general, the effect of its presence in keeping down waves overbalances its effect as a tool of erosion.

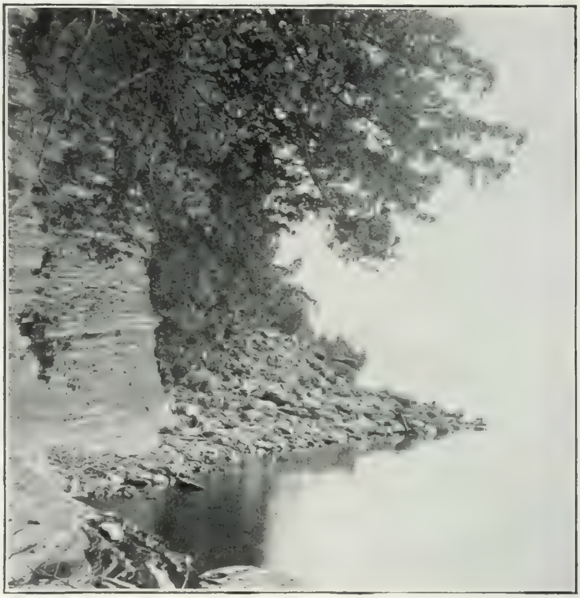


Fig. 220. — Showing blocks similar to those of Fig. 219, reduced and rounded by wave-action. Shore of Lake Champlain. The rock is Utica shale. (Perry.)

The direct effect of wave-erosion is restricted to a narrow zone. There is no impact of breakers at levels lower than the troughs of the waves, though erosion may extend down to the limit of effective agitation. The upper limit of effective wave-action is the level of the wave-crests. The rise and fall of the water during the flow and ebb of the tides gives the waves a greater vertical range than wind-waves alone would have. The indirect work of waves is limited only by the height of the shore, for, as the zone of excavation is carried landward, masses high up the slope are undermined and fall. The fallen rock temporarily protects the shore against the waves (Fig. 219) but the fallen masses are themselves eventually broken up (Fig. 220).

The general result of wave-erosion is the advance of the sea on the land, the rate of advance being determined chiefly by the nature of the material attacked and the strength of the waves. Though examples of the retreat of coast-lines before the advance of seas are numerous, it is not to be understood that the advance of the sea on the land is universal or uninterrupted. On the contrary, the land often encroaches on the sea, and the two things in some places go on side by side. At Long Branch, N. J., the advance of the sea has been so rapid in recent years as to menace important buildings, while a few miles to the north and south, the land is advancing into the sea by the deposition of shore drift. The low coast of the Middle Netherlands has retreated two miles or more in historic times, but the opposite tendency is shown at other points in the same region. On the coast of England, within historic times, the sites of villages have disappeared in the advance of the sea, but other parts of the coast of the same island afford illustrations of land advance. On the south side of Nantucket Island, the sea-cliffs have been known to retreat before the waves as much as six

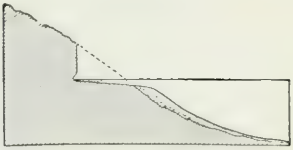


Fig. 221

Fig. 222

Fig. 221. — Diagram illustrating high sea-cliffs. It shows also a submerged terrace, due partly to wave-cutting and partly to building.

Fig. 222. — Diagram showing a low sea-cliff.

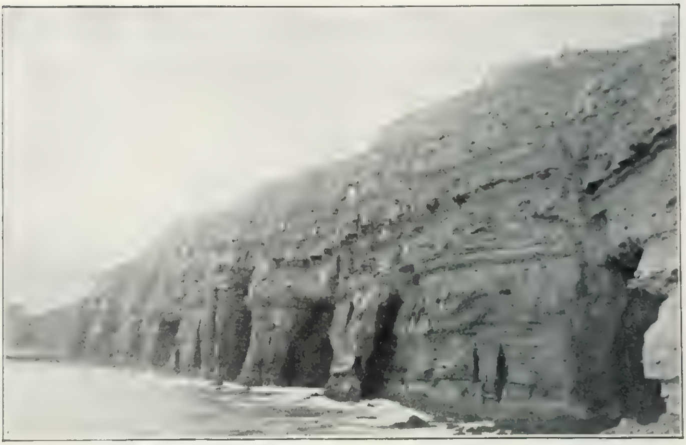


Fig. 223. — A high sea-cliff, La Jolla, Calif.

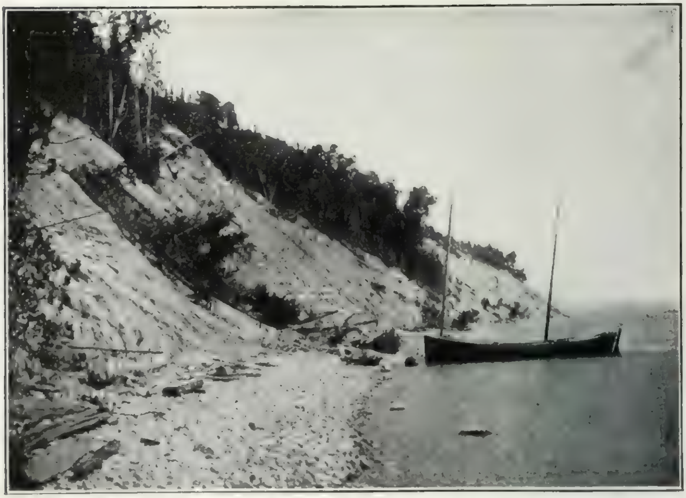


Fig. 224. — A high cliff with a beach, shore of Lake Michigan. (U. S. Geol. Surv.)

feet in a single year. Almost every considerable stretch of coast affords illustrations both of the advance of the sea on the land and of land on the sea; but in the long run, and all coasts considered, the former exceeds the latter.

Topographic Features Developed by Wave-erosion. As the waves cut into the shore at and near the water-level, they develop a steep slope above the line of cutting. This steep slope is the *sea-cliff,* or *wave-cut* cliff (Figs. 221–224). The height of the cliff depends on the height of the land on which the sea is advancing. Its slope may be steep or gentle (Figs. 221 and 222), according to the nature of the material of which it is composed, and the rapidity of cutting. Rapid cutting and resistant material tend to produce steep cliffs; but incoherent materials, such as sand and clay, may form steep cliffs or bluffs if the cutting is rapid. The structure of the cliff-rock also influences the slope and configuration of the sea-cliff.

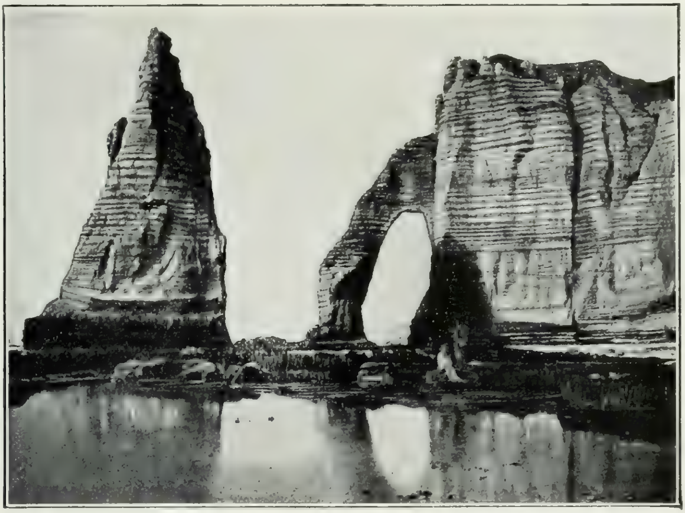


Fig. 225. — A chimney rock and an arch on the coast of France. (Neurdein.)

Waves, working in along the joints of rock, widening them and quarrying out the intervening blocks, isolate pillars of rock ("chimney rocks," "pulpit rocks"), or even considerable islets (Figs. 225 and 226).

Waves in places excavate *caves* at the bases of cliffs. The bottom and roof of a sea-cave usually have a pronounced upward inclination landward, and if the cliff is low, the cave may be extended landward until its roof is pierced. Through such an opening in the top of the cave the water of the incoming waves may be forced in the form of spray. On the New England coast, such holes are sometimes known as "spouting horns." Similar openings may be made, as already pointed out, by the compression or rarefaction of the air in the cave as the wave enters or retreats. The cave, the "spouting horn," the "pulpit rock," and isolated islets, are all closely associated with the sea-cliff in origin.

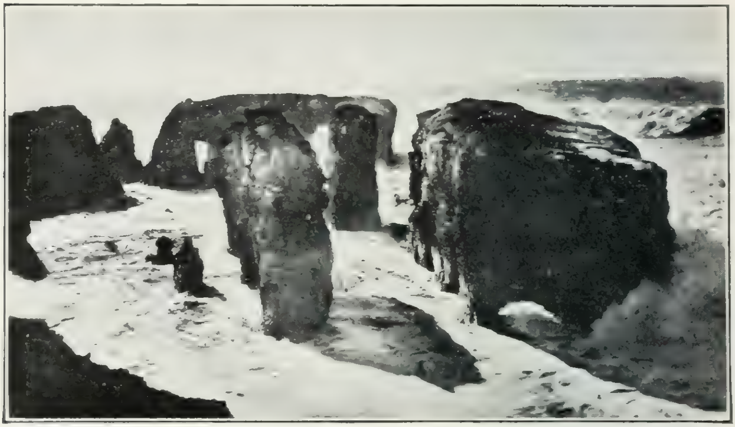


Fig. 226. — Chimney rocks, etc., developed by wave erosion. Coast of California. (Fairbanks.)

The bottom of the sea-cliff is bordered by a submerged platform over which the water is shallow. This platform, or at any rate its landward portion, represents the area over which the water has advanced as the result of wave-cutting, and is therefore known as the *wave-cut terrace*. Such a

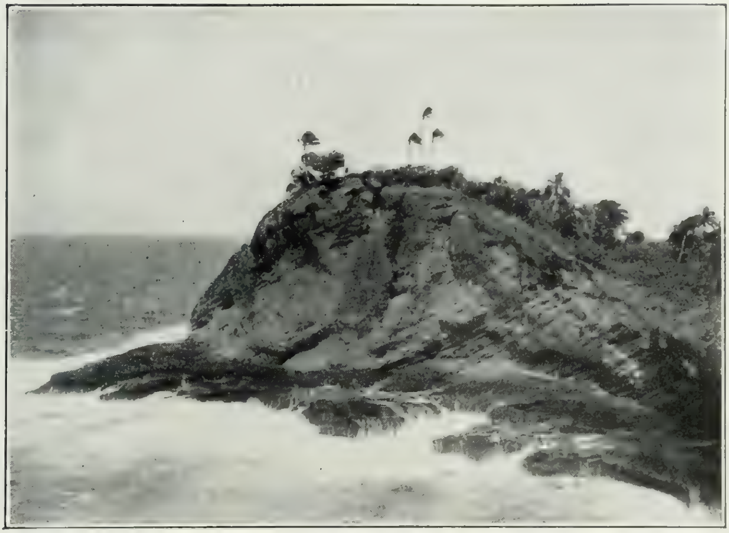


Fig. 227. — Wave-cut terrace on the island of Aunuu, Samoa. A dissected volcanic cone. (R. T. Chamberlin.)

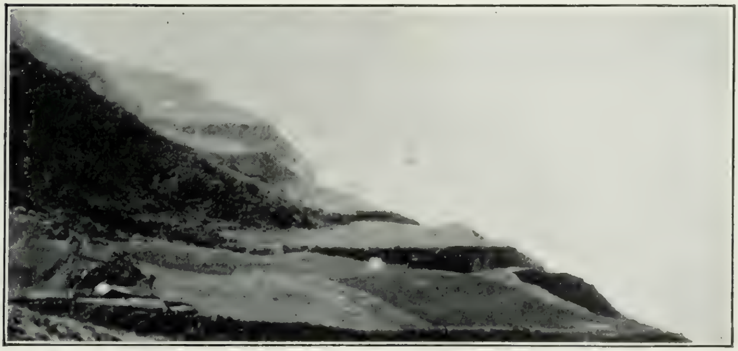


Fig. 228. — Raised sea-terrace, Santa Lucia Mts., Calif. (Courtesy John Burns.)

terrace is the necessary accompaniment of the cliff. Wave-cut terraces may become land by elevation, or by lowering of the level of the sea (Fig. 227). Elevated sea-cliffs with wave-cut terraces at their bases are among the best evidences of relative change of level between water and land (Fig. 228).

Wave-erosion and Horizontal Configuration. The structure of the rock along the shore has much to do with the horizontal configuration of the wave-shaped coast. In general, waves develop re-entrants in the weaker portions of the shore, leaving the more resistant parts as headlands (Fig. 1, Pl. XVIII). It is to be noted that the resistance of rock to wave-erosion is not determined by its hardness alone. Every division plane, whether due to bedding, to jointing, or to irregular fracture, is a source of weakness to the rock, and rock of great hardness may be so broken as to offer relatively little resistance. A coast which is regular and of equal exposure, but of unequally resistant material, will be made irregular by the waves. A regular coast of uniform material, but of unequal exposure, due to protecting reefs or islands, will be made irregular by greater cutting at the points of greater exposure. A coast of marked irregularity and homogeneous material will be made more regular by the cutting off of the projecting points, because they are more exposed. With a given set of conditions, waves tend to develop a certain sort of shore-line which, so far as its horizontal form is concerned, is relatively stable. Such a shore-line may be said to be *mature* so far as wave-erosion is concerned. Since coastal lands are, in general, both heterogeneous and unequally exposed, a mature coast-line is commonly somewhat irregular.

Since the conditions of erosion along coasts are constantly, even if slowly, changing, maturity is constantly being approached, but rarely reached. Other forces and processes, such as those of aggradation, vulcanism, and diastrophism, are in operation along coasts, and their results are sometimes

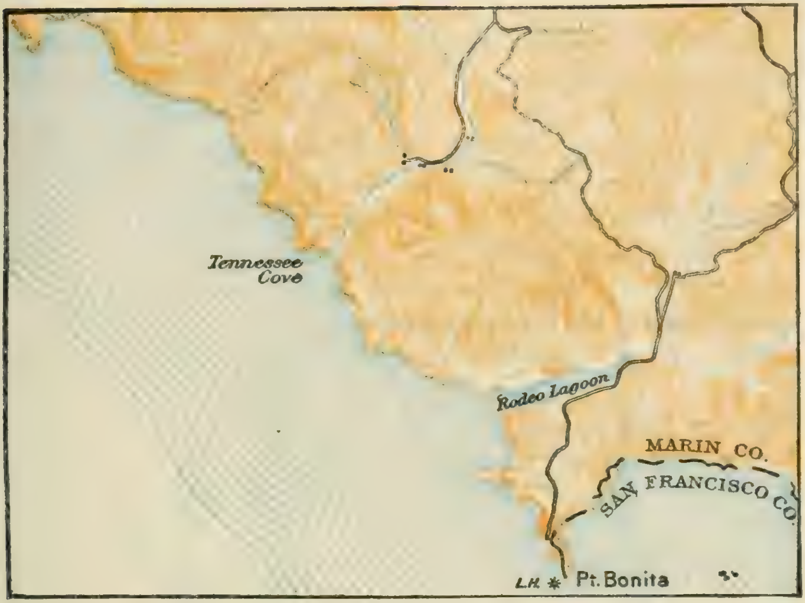

Tennessee Cove
Rodeo Lagoon
MARIN CO.
SAN FRANCISCO CO.
L.H.
Pt. Bonita

FIG. 1.—A coast line developed chiefly by wave erosion. Scale, about 1 mile per inch. (Tamalpais, Cal., Sheet, U. S. Geol. Surv.)

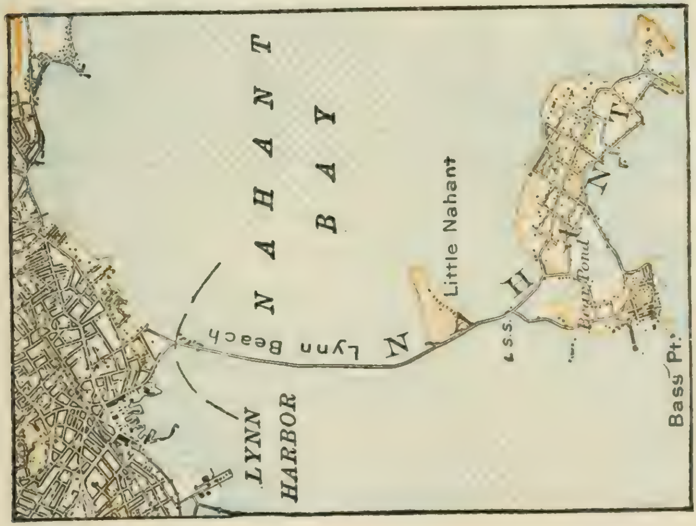
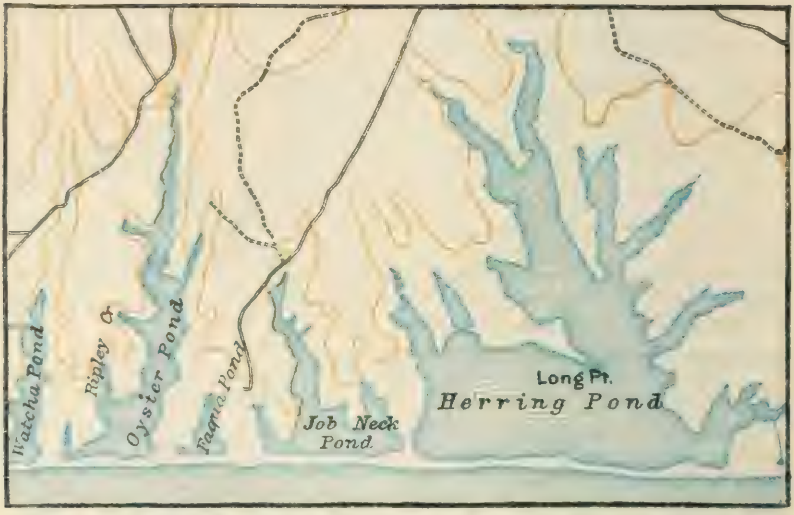

NAHANT BAY
LYNN HARBOR
Lynn Beach
Little Nahant
NAHANT
L.S.S.
Bear Pond
Bass Pt.

FIG. 2.—An island tied to the mainland by a "beach." Scale, about 1 mile per inch. (Boston Bay, Mass., Sheet, U. S. Geol. Surv.)

PLATE XIX


Watcha Pond
Ripley Cr.
Oyster Pond
Fagan Pond
Job Neck Pond
Long Pt.
Herring Pond

Coastal lakes formed by the blocking of the ends of drowned valleys. Scale, about 1 mile per inch. (Marthas Vineyard, Mass.. Sheet, U. S. Geol. Surv) antagonistic to those of the waves. The horizontal configuration of coasts is, therefore, the result of many cooperating forces of which the waves are but one. It is, nevertheless, important to note the end toward which the waves are working, even though they are continually defeated in their attempt to reach it. Their immediate goal is maturity of configuration; their final goal is the destruction of the land and the deposition of its substance in the sea.

Transportation. The material eroded from the shore by the waves is transported by the joint action of waves, undertow, and shore-currents.

The incoming waves of oscillation begin to shift material back and forth on the bottom when the movement of the water particles in the lower part of the waves becomes great enough to handle the material encountered. There is a slight excess of forward movement in the water particles, so that, if the bottom is flat enough, there will be gradual shifting of material shoreward. But if the bottom is steep enough, the movement of material will be seaward. In other words, there is a critical angle of the bottom where equilibrium is established between the shoreward and seaward transportation of material under average conditions. Shoreward from the line of breakers, waves of translation tend to move material landward, while the returning water of the undertow strives to shift it seaward. The resulting transportation of material, whether landward or seaward, depends on the balance of power established between these various factors. This balance varies from place to place and from time to time.

The advance and retreat of waves which come in at right angles to the shore do not move sediment along the shore; but oblique waves and littoral currents do shift it along shore. The direction of movement is readily inferred from Fig. 218. The direction in which debris is shifted by waves and shore-currents is modified by the undertow, and the direction which would result from undertow and current is modified by the waves. It is often the waves of storms rather than those of the prevailing winds which determine the direction of greatest shore transportation.

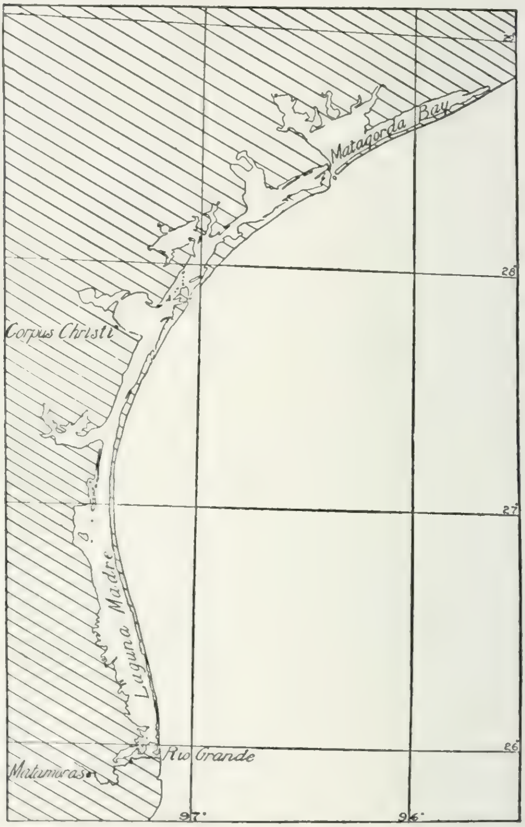

29
Matagorda Bay
28
Corpus Christi
27
Laguna Madre
26
Rio Grande
Matamoras
97°
96°

Fig. 229. — A portion of the Texas coast, showing the tendency of shore deposition to simplify the coast line. The deposits (the narrow necks of land parallel to the coast) shut in the bays. (From chart of C. and G. Surv.)

The waves, the undertow, and the littoral currents work together in assorting the detritus of the shore. If the coarsest parts are beyond the power of all but the strongest waves, they accumulate where the agitation is great. Less coarse parts are carried farther from the site of greatest agitation, but no materials to be classed as coarse are carried beyond the depth of sensible movement. All material which covers the bottom where the agitation of the water is effective, constitutes *shore-drift*, whether derived from the shore by the waves, or brought to the sea by streams.

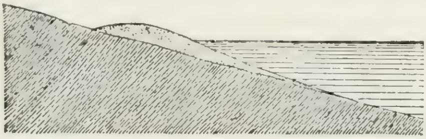


Fig. 230. — Cross-section of a beach. (Gilbert.)

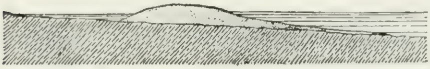


Fig. 231. — Section of a barrier. (Gilbert, U. S. Geol. Surv.)

The material which is fine enough to be held in suspension is measurably independent of depth. This is shown during storms when the water becomes turbid far beyond the line of breakers, and clears only after the waves have died way.

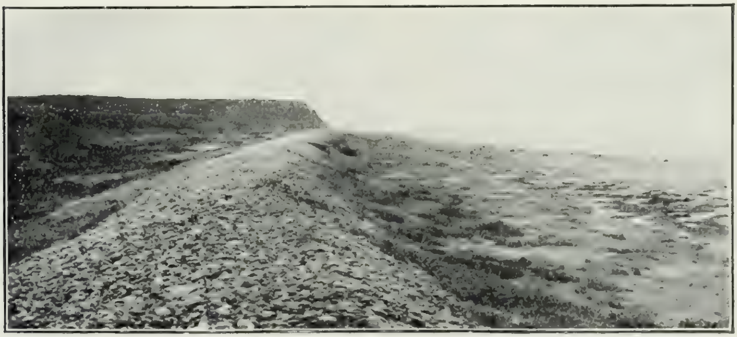


Fig. 232. — An elevated barrier beach-ridge on the coast of California.

The sorting of shore-drift effected while it is in transportation is often very perfect. The conditions favoring assortment are (1) vigorous wave-action, (2) prolonged transportation, and (3) a moderate volume of sediment.

Deposition by Waves, Undertow, and Shore-currents. The zone occupied by shore-drift in transit is the *beach*. Its lower margin is beneath the water, a little beyond the line where the great storm-waves break at low tide. Its upper margin is at the highest level reached by storm-waves, and is usually a few feet above the level of still water during high tide. Material is brought to the beach by incoming waves, and from it detritus is carried away by the undertow. The cross-section of a beach is shown in Figure 230. The beach follows the general boundary between water and land, though it does not conform to its minor irregularities

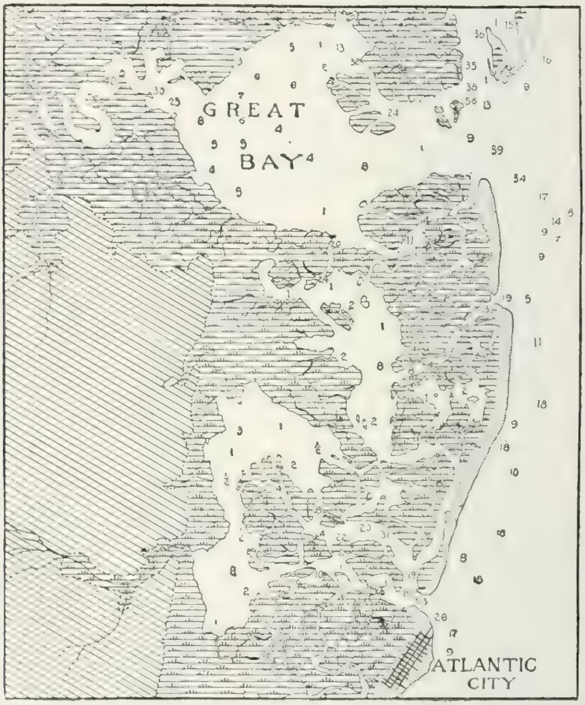


Fig. 233. — Sketch-map of a part of the New Jersey coast. The dotted belt at the east is the barrier, modified by the wind. The area marked by diagonal lines is the mainland; the intervening tract is marsh-land. The numbers show the depth of water in feet. Scale: $\frac{1}{4}$ inch = 1 mile.

(Fig. 229). The beach in places causes the deflection of the lower courses of streams descending to it.

In its deposition the shore-drift assumes various forms. Where the bottom near the shore has a very gentle inclination, the incoming waves break some distance from the shore-line, and it is here that the most violent agitation occurs when the waves are strong. To this line of breakers, material is shifted from both directions. Accumulating here, it builds up a low ridge called the *barrier,* or *off-shore bar* (Fig. 231). If it is built up above the surface of the water by storm-waves, it may shut in a lagoon behind it, and this may ultimately be filled by sediment washed down from the land. At one stage in the filling the lagoon becomes a marsh (Fig. 233).

The deposition of shore deposits depends largely on the currents at and near the shore. If the coast line is deeply indented, the littoral current usually fails to follow the reentrants. In holding its course across the mouth of a small bay, a shore-current usually passes into deeper water. Here its velocity is checked because its motion is communicated to the water beneath, and a larger amount of water being involved in the motion, the motion of each part is diminished. If sediment was being moved along its bottom before the current was checked, some part of it is dropped when and where the current is slackened. It follows that deposition commonly takes place beneath a littoral current as it crosses the mouth of a bay. The belt of deposition is often narrow, and the result is the construction of a ridge beneath the water in the direction of the current. The current would never build the embankment up to the water-level, but when its surface approaches the level of effective agitation, the waves may build it up to, and even above the surface of the water. So long as the outer end of such an embankment is free, it is a *spit.* The construction of a spit has been aptly compared to the construction of a railway embankment across a depression. The material is first carried out from the bordering upland (in this case the shallow water) and dumped where the slope to the depression (deep water) begins. The embankment thus begun is extended by carrying out new material which is left at the end of the dump already made, as at the end of a railway fill.

The spit is normally either straight or parallel with the general course of the shore-current, but since the littoral current is subject to change with the shifting of the winds, the spit may depart from straightness. Winds which simply reverse the direction of the littoral current retard its construction; but waves or cross currents sweeping past the end of a spit, may cause it to curve into a *hook* (Fig. 234). Successive storms may develop successive hooks along the side of a growing spit. The end of a hook may be so extended as to join the mainland, when it becomes a *loop*. If the spit is lengthened until it crosses, or nearly crosses the

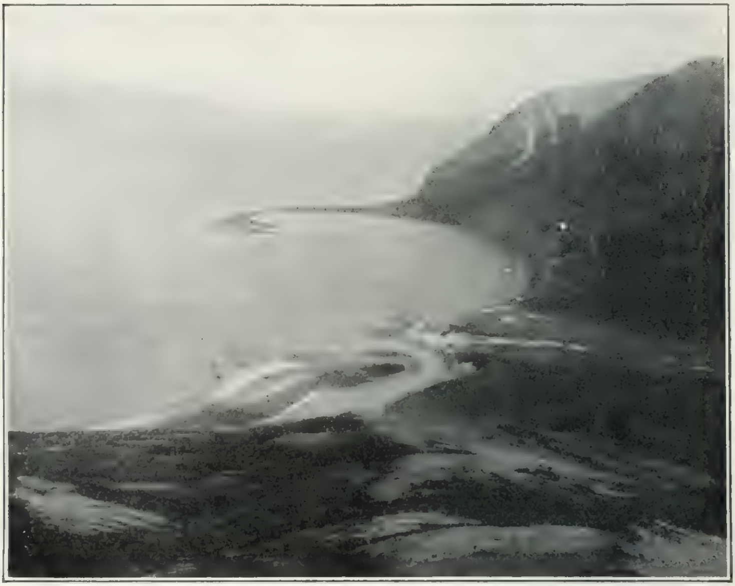


Fig. 234. — Sand and gravel hook at Hook Bay, Alaska. (Atwood U. S. Geol. Surv.)

bay, shutting it off from the open water, it becomes a *bar*. Bars have shut in lakes, ponds, and lagoons at numerous points both on the Atlantic and the Pacific coasts as illustrated by the coast of Martha's Vineyard, Mass. (Pl. XIX). The same phenomena are to be seen along many lake shores (Fig. 235). Bars in places tie islands to the mainland (Fig. 2, Pl. XVIII), in which case they are called *tombolos* (Fig. 236). If the bay across which the bar is built receives abundant drainage from the land, the outflow from the bay

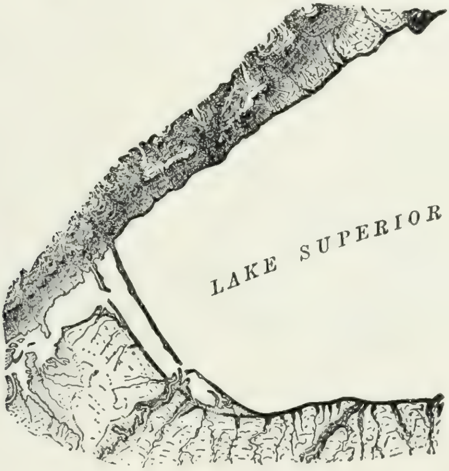


Fig. 235. — Map of the head of Lake Superior. (U. S. Geol. Surv.)

may be sufficient to prevent the completion of the bar, for when the growth of the bar has sufficiently narrowed the outlet of the bay, the sediment brought to the end of the spit by the littoral current will be swept out beyond the spit by the current setting out from the bay. The completion of a bar may be interfered with also by tidal currents, even without land-drainage. The scour of the tides in many places preserves deep entrances (inlets) to bays, and maintains definite channels or "thorofares" in the lagoon marshes behind barriers and spits. The sediment brought down from the land, as well as that washed in by tidal currents and waves, tends to fill up the lagoon behind a barrier, a spit, or a bar, converting it into land (Fig. 233).

The off-shore movements of shore-waters may leave the sediment of the shore in the form of a *wave-built terrace,* which is really a seaward extension of the beach. The wave-built terrace commonly borders the seaward margin of the wave-cut terrace (Fig. 221). Terrace-cutting and terrace-building are both involved in the development of the continental shelves. Beach-ridges, spits, bars, etc., like sea-cliffs and wave-cut terraces, may be preserved after the relative level of the sea and land has changed. If the sea-shore has risen, relatively or absolutely, these features are evidence of the change. If shore features are submerged instead of elevated, they furnish less accessible, though not less real, evidence of the change of level. Similar features about lakes have a like significance, but in this case it is often demonstrable that it is the water rather than the land which has changed its level.

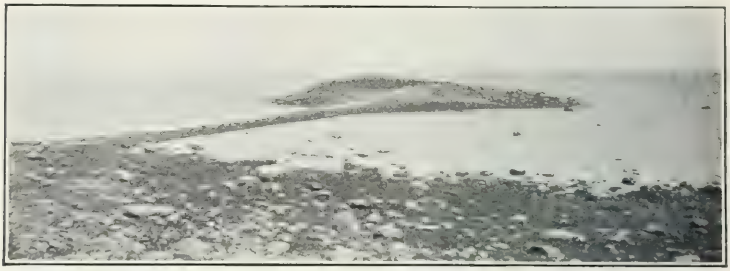


Fig. 236. — Tombolo, Sheep Island. Penobscot Bay, Me. (Bastin, U. S. Geol. Surv.)

EFFECT OF SHORE-DEPOSITION ON COASTAL CONFIGURATION

The tendency of shore-deposition is to cut off bays and to straighten and simplify the shore-lines. This is abundantly illustrated along the Atlantic and Gulf coasts of the United States (Fig. 229 and Pl. XIX). It is to be noted, however, that in the simplification of the shore-line through deposition, the initial stages often result in great irregularity (Fig. 2, Pl. XVIII).

Currents

Two types of currents are found in the ocean. One is the more or less localized movement of the water at or near the surface, while the other type is the slow, continuous movement, "*creep*," of the deep-sea and bottom water.

Since surface currents, which if wide and indistinct are called *drifts*, move in the general direction of the atmospheric circulation, it may well be inferred that they are generated primarily by the wind. This is particularly true in tropical latitudes. After the currents are generated, their directions may be controlled by such other factors as the configuration of the continents or ocean bottoms, by the rotation of the earth, and by the winds in the region into which they move.

As agents of erosion, ocean-currents are not of great importance. Currents which reach the bottom are comparable, in their effects, to rivers of the same velocity and volume; but most ocean-currents do not touch bottom, and therefore do not erode it. Only where they flow through narrow and shallow passageways is their abrasive work considerable. Thus the Gulf Stream has a velocity of four or five miles per hour where it issues from the Gulf of Mexico, and its shallow and narrow channel is current-swept. Other illustrations of the erosive power of currents have been noted in analogous situations, but in spite of the known examples, it yet remains true that ocean-currents, on the whole, are but feeble agents of erosion.

Ocean-currents are scarcely more important in transporting and depositing sediment, than in eroding. Currents which do not touch bottom roll no sediment, and carry only what may be held in suspension. If a particle of mud in suspension in a river drops to the bottom, as it frequently does, it may be picked up again and carried forward. But if a particle in suspension in an ocean-current once escapes the moving water by settling through it, the current which does not drag bottom has no chance to pick it up again. Very fine sediment (however acquired) may be carried by an ocean-current far beyond the place where it was acquired, but currents which do not touch bottom are rarely strong enough to hold any but the finest material for any considerable length of time. However, small plants and animals which inhabit particularly the warm waters in great numbers are carried far and their remains may be widely distributed.

But as modifiers of climate, ocean currents are of great importance. The winter climate of Europe is much less severe than it would be without the Gulf Stream (North Atlantic drift); and the northwestern part of North America is warmed by the Japan Current in the North Pacific. The Humboldt (Antarctic) Current, on the other hand, keeps the west coast of South America much cooler than it otherwise would be.

The deep-sea circulation today is dominated by the slow movement of cold water from the poles toward the equator. This circulation is undoubtedly due, in large part, to the sinking, in polar latitudes, of heavy water, produced by great cooling and by increased salinity of the unfrozen waters during times of surface freezing. Likewise movements due to increased salinity in the belts of great evaporation doubtless are factors of importance in the deeper circulation.

DEPOSITS ON THE OCEAN-BED

The deposits on the bed of the ocean may be divided into two classes according to the depth of water in which they are made. *Shallow-water deposits* are made in water less than about 100 fathoms deep, and *deep-sea deposits* are laid in water of greater depth. The selection of the 100-fathom line as the dividing depth is less arbitrary than it seems, for passing outward from the shore, it is at about this depth that the bottom commonly ceases to be disturbed by the action of currents and waves, that sunlight and vegetable life cease to be important at the bottom, and that the coarser sediments which predominate along shore give place to muds and oozes. It is important to have in mind also that the 100-fathom line (or some line very near it) is an important one in the physical relief of the globe, for it appears to mark, approximately, the junction of continental platforms and ocean-basins. Because the latter are a little over full, the water runs over their rims, covering about 10,000,000 square miles of the borders of the continental protuberances.

Aside from the deposits made by organisms, shallow-water deposits are divisible into two groups — (1) those immediately along the shore, the *littoral deposits*, and (2) those made between the littoral zone and the 100-fathom line. Both are chiefly *terrigenous* (land-derived), though chemical and organic deposits occur in both zones. The deep-sea deposits likewise are divisible into two principal groups — (1) the *terrigenous deposits* near the land, and (2) the *pelagic deposits*, made up chiefly of the remains of pelagic (free floating) organisms and the ultimate products arising from the decomposition of rocks and minerals. The former predominate in the deep waters relatively near the shore; the latter in the deeper water far from land. The shallow and deep-water deposits grade into each other in a belt along the 100-fathom line.

SHALLOW-WATER DEPOSITS

Littoral Deposits. The littoral zone is often defined as the zone between high- and low-water marks, but in common usage, the very shallow water a little farther from the coast-line is often included. It is the zone in which boulders, gravels, sands, and all the coarser materials accumulate, though muds are occasionally met with in sheltered estuaries. Generally speaking, the nature of these deposits is determined by the character of the adjoining lands and the nature of the local organisms. The heavier materials brought down by rivers or worn from the shore by waves are here spread out by waves and shore-currents. Twice in the twenty-four hours the littoral zone is covered by water, and twice parts of it are exposed to the direct rays of the sun or the cooling effects of the night. Physical conditions in general are here most varied. Still greater diversity is introduced by the fact that the zone is inhabited by both marine and terrestrial organisms, while the evaporation of the sea-water which flows over tidal marshes and lagoons may lead to the formation of saline deposits. The length of the coast-lines of the world is some 125,000 miles, so that the zone of littoral deposits, though narrow, covers a very considerable area.

MECHANICAL DEPOSITION IN SHALLOW WATER BEYOND THE LITTORAL ZONE

These deposits are made between the littoral zone and the 100-fathom line, and cover an area of about 10,000,000 square miles, which is greater than the area of the entire North American continent. Their composition is much the same as that of the littoral deposits with which they are continuous, except that they are, in general, finer. At their seaward limit they pass insensibly into the fine deposits of the deep sea. Coarse materials, such as gravel and sand, are most abundant near coasts where such material is supplied, and where agitation is strong enough to transport it. Elsewhere, as along low shores, etc., and also out toward the oceanward edge of the zone, muddy deposits dominate. Some of the deposits are composed almost wholly of inorganic debris, but organic remains are freely mingled with others. The mechanical effects of tides, currents, and waves are everywhere present, but become less and less well marked as the 100-fathom line is approached. The forms of vegetable and animal life are numerous, though the former decrease as depths which make the sunlight feeble are approached.

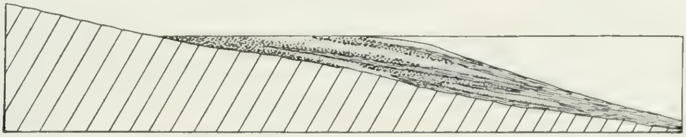


Fig. 237. — Diagram showing the interwedging of gravel, sand, and mud beds.

As a rule, no definite line marks the seaward terminus of the coarse detritus, since coarse material is carried farther out when the waves run high (and undertow is strong) than when they are feeble. In calm weather, therefore, the fine sediment may be deposited where coarse had been laid down in the preceding storm, only to be covered in turn by other deposits of a different character. Thus gravel grades off into sand, with more or less overlapping or interwedging,

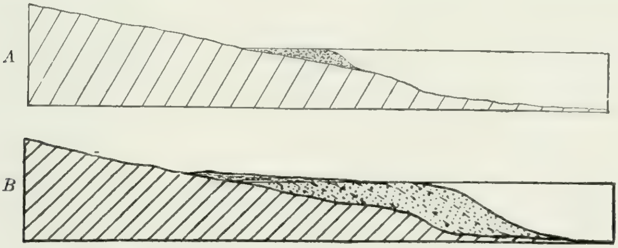


Fig. 238. — Diagrams showing how shallow-water deposits may attain considerable depth by the shifting of the zone of deposition seaward.

and sand grades off into silt in the same way. This is diagrammatically illustrated in Figure 237.

Since coarse deposits may extend far out from land where the waves are strong and the water shallow, and since the zone of shallow water may be extended seaward by the aggradation of the bottom, shallow-water deposits may cover extensive areas. They may become thick at the same time, for as the outer border of the shallow-water zone is shifted seaward by aggradation, the vertical space to be filled becomes greater (compare *A* and *B*, Fig. 238). Again, if the coast is sinking, new deposits of coarse material may be made on older ones. In this way, also, great thicknesses of sediment may be accumulated, all parts of which were deposited in shallow water. The great thicknesses of some of the shallow-water marine formations of the past show how far this process may go.

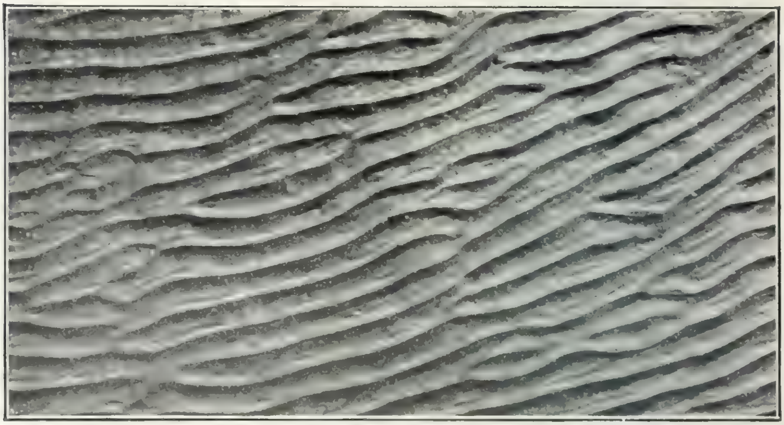


Fig. 239. — Ripple-marks.

Characteristics of Shallow-water Deposits. Fragmental (clastic) sediments laid down in shallow water have several distinctive characteristics. While they are, in the aggregate, coarse, they are characterized by frequent variations in coarseness. The surfaces of successive beds here and there are ripple-marked and in rarer instances rill-marked (Figs. 239 and 240), while cross-bedding (Fig. 241) is very common. Clayey sediments, at times uncovered by water and exposed to drying, are likely to show mud-cracks (Fig. 243), and the tracks of land animals are sometimes found preserved on their surfaces. Shallow-water deposits commonly contain

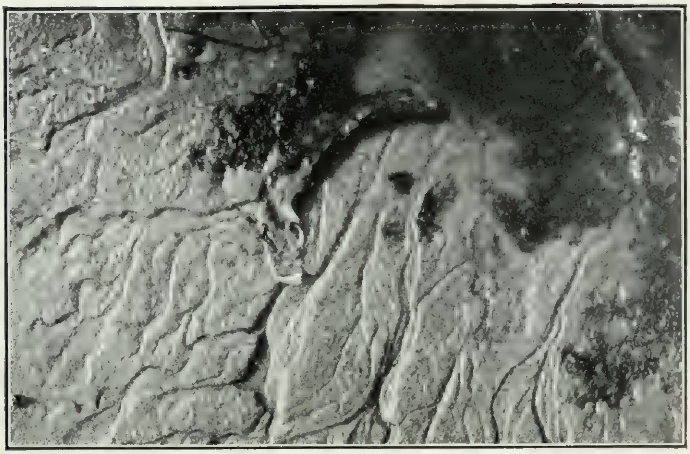


Fig. 240. — Rill-marks resembling impressions of seaweeds. Beach at Noyes Point, R. I. (Walcott, U. S. Geol. Surv.)

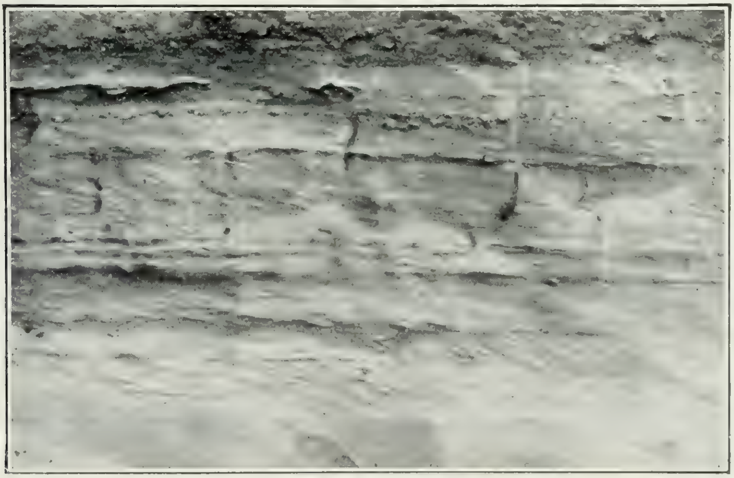


Fig. 241. — Cross-bedding.

fossils of organisms, the majority of which lived in waters of slight depth. These characteristics are sufficient to distinguish sedimentary formations made in shallow water, even after they have been converted into solid rock, and after the rock has emerged from the sea. Many of these characteristics are, however, shared by deposits made by streams on land. Sub-aerial and lacustrine sediments are usually distinguished from those made in the sea by their fossils, and sometimes by their distribution and structure.

Topography of Shallow-water Deposits. The shallow-water deposits have, on the whole, a rather plane surface, though there are some notable departures from flatness. The steep slopes of delta-fronts and of wave-built terraces have already been spoken of. Barriers often shut in depressions, and the disposition of the material is sometimes uneven, owing to shore and tidal currents. The result is that the surface of the shallow-water deposits is not uncommonly affected by low elevations and shallow depressions. The elevations and depressions may be elongate, circular, or irregular in form. These slight topographic irregularities give rise to cross-bedding and other departures from horizontal stratification (Fig. 242).

NON-CLASTIC DEPOSITS IN SHALLOW WATER

Limestone. Lime-secreting organisms inhabit the shallow waters of the oceans in great profusion. They extract the calcium carbonate from the sea-water to build shells. coral reefs, bones, teeth and other devices for supporting, housing, and protecting themselves. When the organism dies and decays these hard parts remain and accumulate slowly upon the bottom. Effective agitation of waves may grind them into lime mud. Certain groups of plants (algae) are found to cause the deposition of lime in considerable quantity. Bacteria, in consequence of their chemical activities, are known to precipitate large quantities of lime in some of the warmer parts of the ocean. As shown earlier, calcium is

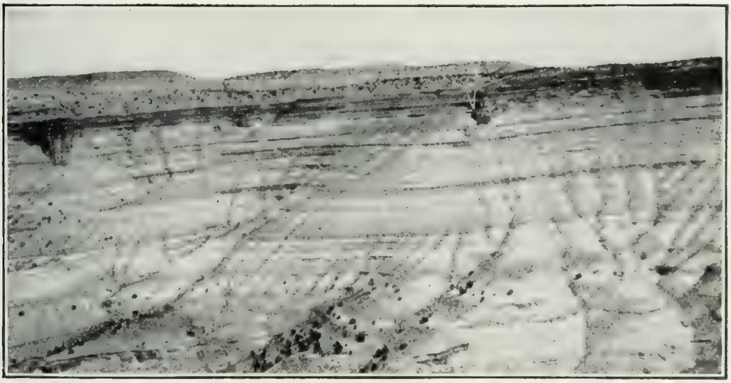


Fig. 242. — Horizontal stratification of shale, Utah. (U. S. Geol. Surv.)

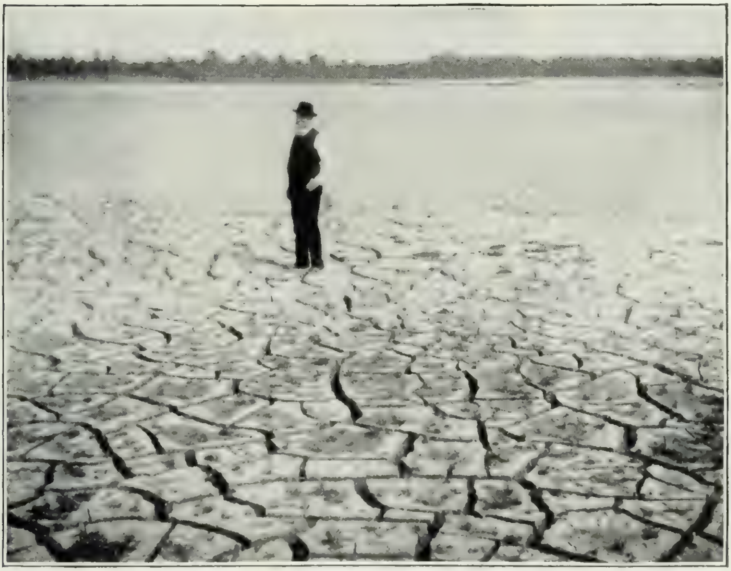


Fig. 243. — Mud-cracks. Flat of the Missouri River a few miles above Kansas City. (R. T. Chamberlin.)

brought by rivers to the sea in the form of soluble calcium bicarbonate ($Ca(HCO_3)_2$). In the sea, part of the carbon dioxide may escape, causing part of the less soluble calcium carbonate ($CaCO_3$) to be precipitated. Calcium carbonate is also precipitated from sea-water in several other ways. This chemically precipitated calcium carbonate mixes with the organic lime mud accumulating on the ocean-bottom to form layers which, upon solidification, become *limestone.*

Since lime-secreting organisms inhabit all epicontinental seas, calcium carbonate is being deposited at all times on the continental shelves. It is obvious, however, that this material accumulates very slowly, so that if gravel, sand, or mud is being brought by rivers, waves, or currents faster than the calcium carbonate is deposited, a conglomerate, sandstone, or shale, containing some calcareous material, will eventually result. It is evident that limestone will be deposited wherever clastic sediments are either not present, or are being deposited in very small amounts. Such conditions are found in clear water, either far from shore, or in favorable places where land drainage, waves, or currents bring little or no clastic sediment.

Other Chemical Deposits. *Gypsum* (the second substance to be precipitated when sea-water is evaporated) appears to be deposited in quantity only where, through evaporation, concentration of the salts in solution reaches an advanced state. Since normal sea-water is far from saturated with *common salt* (the third precipitate to form upon the evaporation of sea-water), the latter is precipitated only in lagoons, in closed areas, or other situations favorable to great concentration. This is, as a rule, only in regions which are notably arid. It follows that deposits of salt usually signify highly arid conditions, and where they occur over wide ranges in latitude and longitude, as in the rocks of certain periods of the past, general aridity of climate is inferred. Where confined to limited areas, their climatic significance is less, for topographic conditions may determine local aridity. The total area where salt is now being precipitated is small, though on the whole the present is probably a rather arid period of the earth's history. On the other hand, ancient deposits of salt preserved in the sedimentary strata show that the area of salt deposition has been much more considerable than now at one time and another in the earth's history. The salt and gypsum deposits of the past seem, therefore, to tell an interesting tale of the ancient climates of the earth.

The *magnesium salts* are among the last to to be thrown down as sea-water is evaporated, being for the most part in the form of sulphates and chlorides. In the artificial evaporation of sea-water to obtain common salt, the process is usually stopped before the saturation point for the magnesium salts is reached, and the residue, the "mother-liquor," or "bittern," is drawn off to prevent these "bitter" salts from mixing with the common salt. The magnesium salts are among the last to be precipitated, not only because they are readily soluble, but because their quantity is small; yet in the original rock from which the sea-salts came, there is nearly as much magnesium as sodium, while in the sea there is about five times as much sodium as magnesium. This is because much of the magnesium from the sea-water has entered into the formation of *dolomite* (magnesian limestone), which is relatively insoluble. In the earlier marine strata, dolomite, composed partly or wholly of the double carbonate of calcium and magnesium $(CaMg)CO_3$, is very abundant. This dolomite appears to have been formed by the gradual substitution of magnesium for a part of the calcium, in the calcium carbonate deposited on the floor of the sea.

Certain plants and animals of the sea secrete silica in considerable quantity. Deposits of this sort are relatively more important in the deep sea than in shallow water, and will be mentioned in that connection.

DEEP-SEA DEPOSITS

Contrasted with Shallow-water Deposits. The deep-sea deposits cover the ocean-bottom below the 100-fathom line. Their area is considerably more than half the earth's surface. The characteristic deposits are muds, organic oozes, and clays, which, in their physical characteristics, are remarkably uniform. Gravels and sands are rarely found. In regions of floating ice, some diversity is introduced from the varied nature of the materials which ice transports. Vegetable life is limited to the surface waters, while animal life is present both in them and on the bottom, but absent or nearly so in the great middle zone between. The temperature of the bottom is at present below 40° Fahrenheit throughout the larger part of the ocean, and subject to little variation. The conditions over the sea-bottom are therefore very uniform.

In consequence of the slow accumulation of sediment on the deep-sea bottom, the absence of transportation there, and the nature and size of the particles, many evident chemical reactions take place, resulting in the formation of secondary products such as glauconite, phosphate and manganese nodules, zeolites, etc. The amount of these secondary products increases, relatively, with increase of distance from land. At the same time there is an increase (relative) of the remains of pelagic organisms with increasing depth. We thus pass insensibly from deep-sea deposits of a terrestrial origin (terrigenous deposits) near the land, to *pelagic deposits*, more or less free from land-derived material.

Mechanical Deposits. The mechanical deposits of the deep sea come 1) from the land, by the ordinary processes of gradation, (2) from volcanic vents, and (3) from extra-terrestrial sources. The terrigenous materials which reach the deep sea are, as a rule, only the finest products of land decay and are carried out by movements of water and by winds. They are not commonly recognized in the dredgings more than 200 miles from the shore, but opposite the mouths of great rivers they extend much farther, — 1000 miles in the case of the Amazon. They are especially abundant on the slopes of the continental shelves, where the *blue*, *green*, and *red muds* are associated with volcanic and coral muds. The color of these various muds depends, in part, on the changes they have undergone since their deposition. The green muds usually contain enough glauconite to give them their color, and are most commonly found off coasts where sedimentation is not rapid The blue muds indicate lack of oxidation, or perhaps reduction of the iron compounds. Red muds are found particularly where tropical rivers have worked red soils into the ocean.

The occasional presence (in the deep-sea deposits) of coarse materials derived from the land must be looked upon as in some sense accidental. Pebbles, or even boulders, entangled in the roots of floating trees may be carried out into the ocean, and icebergs carry out boulders and smaller fragments of rock. The amount which icebergs could carry, if fully loaded, is far greater than the amount which they actually do carry.

Of the identifiable inorganic materials among the pelagic deposits, those of volcanic origin are most abundant, and among these the most common is pumice, which often floats widely before it becomes waterlogged. Pieces of pumice brought up from the bottom and thoroughly dried, will float for months in sea-water. The next most abundant substance of volcanic origin in pelagic deposits is *volcanic glass*. This ranges from walnut-sized pieces down to the smallest fragments, which often serve as centers for concretions. *Lapilli* (cinders) and *volcanic dust* also abound in parts of the deep sea. The distribution of these volcanic products is essentially universal, though by no means uniform. Some of them are probably from submarine volcanoes.

The deep-sea deposits contain many nodules and grains which are believed to be of extra-terrestrial origin. Many of them are magnetic. The dust of the countless meteors which enter the atmosphere daily settles on the land and sea alike, and must enter into the sediment at the bottom of the latter. It is probably no more abundant in deep water than in shallow, but relatively more important there, since other sedimentation is meager. The number of meteors which enter the atmosphere daily has been estimated at from 15,000,000 to 20,000,000. If, on the average, the meteorites weigh ten grains each, probably a rather high estimate, the total amount of the extra-terrestrial matter reaching the earth yearly would be 5000 to 7000 tons, and something like three-fourths of this must, on the average, fall into the sea. But even at this rate it would take some fifty billion years to cover the sea-bottom with a layer one foot in thickness. However, there are reasons, which will appear later, for believing that the infall of such matter was greater in the past.

Organic Deposits. With increasing distance from shore and especially with increasing depth of water, terrigenous deposits become less and less abundant, and sediments derived from pelagic life increase in relative importance. Beyond the upper part of the outer slopes of the continental shelves, the deep-sea deposits are made up largely of the secretions of marine organisms which live in the surface waters. When the organisms die, they sink to the bottom and their remains are mingled with the volcanic and other materials which are universal over the sea-floor.

Deep-sea deposits of organic origin are named according to their characteristic constituents. Thus there are *pteropod oozes, globigerina oozes, diatomaceous oozes, radiolarian oozes,* etc. It is not to be understood that these oozes are made up exclusively of the remains of the organisms which give them their names. Diatomaceous ooze is an ooze in which the siliceous secretions of diatoms are abundant, not an ooze made up wholly of them; and likewise globigerina ooze is an ooze in which the globigerina shells are abundant, though in many cases they do not make up even the bulk of the matter. While samples of these various oozes might be selected which are thoroughly distinct from one another, there are all gradations between them, since pelagic life does not recognize boundary lines.

It is a significant fact that with increasing depth the portion of lime carbonate in the ooze decreases. In tropical regions remote from land, where the depths are less than 600 fathoms, the calcium carbonate of the shells of pelagic organisms may constitute 80 or 90 % of the deposit. With the same surface conditions, but with increasing depth, the percentage of calcium carbonate decreases, until at 2000 fathoms it is less than 60 %; at 2400 fathoms, 30 %, and at 2600 fathoms, 10 %. Beyond this depth there are usually no more than traces of carbonate of lime. When the percentage of lime carbonate becomes very low, the calcareous oozes grade off into the *red clay* with which the sea-floor below 2400 to 2600 fathoms is covered.

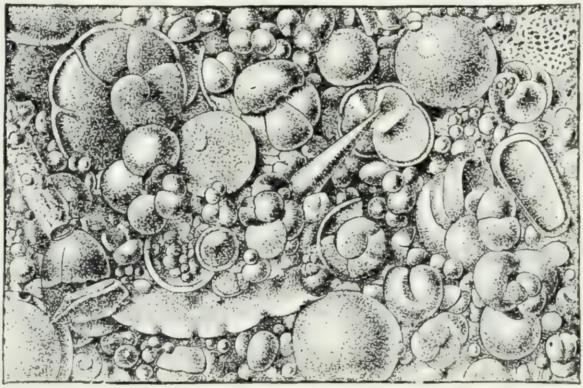


Fig. 244. — Globigerina ooze. Magnified 20 times. (Murray and Renard.)

Chemical Deposits. The chemical deposits of the deep sea are chiefly the alteration products of sediments already on the sea-bottom. Where sedimentation is rapid and the sediment coarse, the chemical change is relatively slight;

but where the sedimentation is slow and the sediment fine, the chemical change is relatively greater, for both the longer exposure to the sea-water and the greater surface of the particles exposed to attack, favor change.

The *red clay* already referred to belongs to this class of deposits. Its origin has been the subject of much discussion. It contains much volcanic debris, various concretions, bones of mammals, zeolite crystals, extra-terrestrial spherules, and doubtless the insoluble parts of the shells of pelagic life; but it is still a mooted question how far the clay itself is a residual product of decomposed shells, and how far the altered product of pulverized pumice, volcanic ash, dust, etc. Pelagic life does not seem to be less abundant at the surface where the water is deep than where it is shallow, and the remains of organisms must sink in such situations as elsewhere. If the lime carbonate of globigerina ooze is removed by dilute acid, the inorganic residue is similar to the red clay in the ocean bottom. This suggests that, owing to the more complete solution in the very deep water, the lime carbonate of the shells has been dissolved, leaving as a residium material which makes up a part of the red clay. The more complete solution at the bottom might be the result either of the greater pressure, or of a greater percentage of CO_2 in the deeper waters. The study of the dredgings, however, has inclined the students of these materials to the conclusion that volcanic debris, rather than shells, is the principal source of the red clay. The volcanic materials are thought to have accumulated slowly, and to have been long exposed to the action of sea-water. The various nodules and crystals in the clay are believed to be secondary products, the materials for which were derived from the decomposition of the sea sediments. Eolian dust, or the material derived from it by chemical alteration, is doubtless a constituent of the red clay.

It is a significant fact that deposits corresponding to those of the deep sea have not been identified with certainty among the rock formations of the land. If such deposits are really absent from the land, as they seem to be, their absence must mean that the continents have never been under the deep seas. That large parts of them have been beneath shallow sea-water is abundantly attested.

CORAL REEFS AND CORAL ISLANDS

In the warmer portions of the oceans where other conditions are favorable, corals live in great profusion. The individual coral animals, or polyps, extract carbonate of lime from the sea-water and in time build up stony bases upon which they live. While there are many kinds of corals, the most important geologically are the reef-building corals which live together in colonies and so develop a composite stony base for the whole colony. As the coral colony grows upward and outward, the living polyps grow steadily away from the older portions of the basal limy structure which remain as ever-enlarging foundations for the colony. Where many branching and spreading colonies start close together, they soon coalesce, and the open spaces gradually become filled with fragments of coral broken off by the waves, together with shells and the solid remains of other marine organisms. Lime precipitated from the sea-water in various ways cements the composite mass together, finally producing a *coral reef*. Normally such a reef is made up of coral masses of many different kinds (Fig. 245) together with the deposits of certain lime-secreting plants (nullipores, etc.) which serve to bind the masses together.

Reef-building corals are so sensitive that they thrive only under certain conditions: (1) The temperature of the water must not fall below 68° F.; (2) the water must not be more than 120 feet deep; (3) it must be of the normal ocean salinity; (4) it must be relatively free from mud or other clastic sediment, and (5) it must be sufficiently agitated to bring adequate food supply. Too low a temperature, too much freshening of the sea-water by streams from the land, too much fine sediment, or a subsidence which carries the reef to too great depths, kills the delicate polyps. The most rapid growth of reef-building corals is just below the surface of the sea where the surging waves bring a constant supply of food.

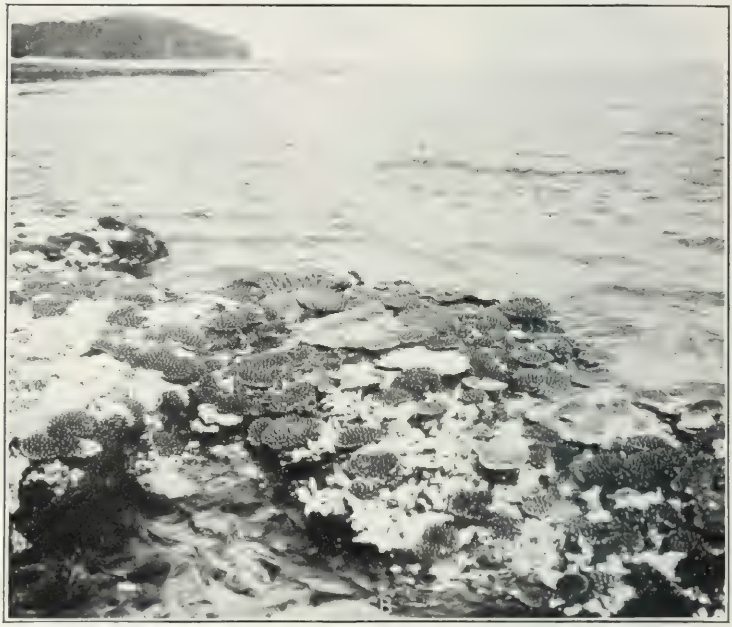


Fig. 245. — Seaward edge of Aua (fringing) Reef at low spring-tide. Tutuila, Samoa. (A. G. Mayor, Carnegie Institution.)

Types of Coral Reefs. The coalescence of many coral colonies along shores in clear, tropical seas develops shore-reefs, or *fringing reefs*. Their seaward margins are normally steep, and may even overhang, for the most active coral growth is on their outer margins where the water is best aërated and the breaking waves supply the most food and lime. If the water is not too deep, coral debris broken off by the waves accumulates beyond the reefs, and the corals, later growing upon this accumulation, may build reefs far out from the shores. During the lowest spring-tides, large patches of a vigorously developing fringing reef commonly become visible above the ocean surface (Fig. 245). Fringing reefs may border shores for many miles (Fig. 246) though interrupted here and there by narrow channels, or gaps, where streams bring in sufficient mud and fresh water to prevent the coral growth.

If a reef forms near the outer edge of a broad, wave-made platform so as to enclose a lagoon of deeper water between it and the shore, it constitutes a *barrier reef*. Barrier reefs, however, develop in other ways, as will be noted presently. Some are of large extent, the largest known being the Great Barrier Reef of Australia which, lying 20–30 miles off-shore, parallels the northeast coast of that continent for at least 1200 miles. The enclosed lagoon has a depth of 100–300 feet.

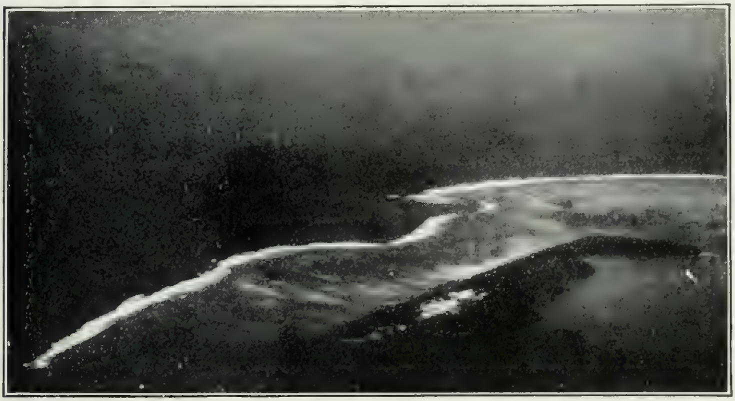


Fig. 246. — Line of breakers on outer border of fringing coral reef around Cocoanut Point, Tutuila, Samoa. At low tide there is very little water upon the reef. From top of Matafao Peak, 2141 feet above the sea. (R. T. Chamberlin. Carnegie Institution.)

If a volcanic island surrounded by a barrier reef becomes slowly submerged (either by the island sinking or by sea-level rising), while corals keep building the reef upward as fast as the island sinks, an *atoll* results when the last of the original island disappears beneath the waves. Atolls are ring-like reefs enclosing relatively shallow lagoons above which no land of any consequence appears (Fig. 247). While corals build the atoll rim only slightly above lowest tide-level, the ocean waves pile up coral detritus and build a chain of islands upon the reef. Once considerable sections of the reef become land, the wind may make them higher by piling up coral sand. Cocoanut groves and other vegetation help along the building, both by their own growth and by helping the lodgment of wind-blown material. The lagoons of most atolls remain connected with the ocean by shallow inlets.

Atolls are particularly abundant and striking in the Pacific and Indian oceans, and are of many sizes. One of the smallest, Rose Atoll in the Samoan Group of the South Pacific, is a ring about 500 yards wide surrounding a lagoon about 2 miles in diameter, while Suadiva Atoll (Maldive Group) in the Indian Ocean is 44 miles north and south and 34 miles across, and the Great Chagos Bank is apparently a submerged atoll 90 miles long and 70 miles in its greatest breadth.

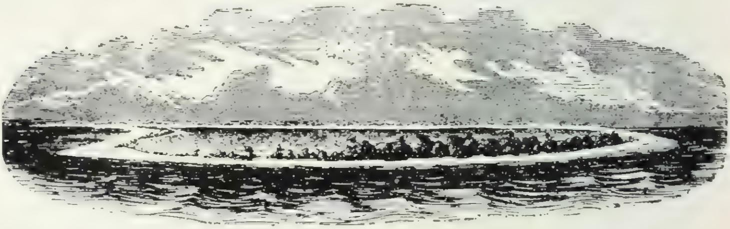


Fig. 247. — An atoll. (From Dana's *Corals and Coral Islands.*)

The development of fringing reefs is easily understood, but the origin of barrier reefs and atolls has long been a debated question. The time-honored explanation has been the theory of subsidence originated by Darwin. According to this theory, corals attaching themselves to volcanic islands in the tropical seas first develop fringing reefs along the shores. With the passage of time, the volcanic pile slowly sinks, but as it sinks the upward growth of coral keeps pace with the subsidence, so that the growing coral-heads maintain themselves near sea-level. The coral growth is more rapid on the outer side of the reef than on the inner side where the restraining influence of land-derived sediment is most felt. The result is that as the volcanic mass sinks and the corals build upward, the fringing reef becomes a barrier reef, separated from the shore by a lagoon. Waves attack the exposed portion of the reef and the detritus is washed down the slope. If the volcano continues to sink until the central land-mass finally disappears beneath the waves, the barrier reef becomes an atoll.

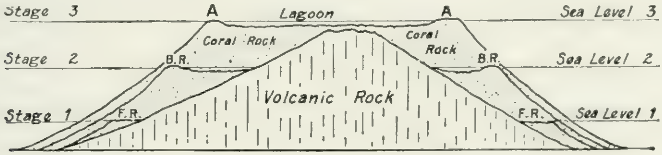


Fig. 248. — Diagram illustrating development of a barrier reef and an atoll, successively, from a fringing reef by sinking. F.R., fringing reef formed along shore at time of sea-level 1; B.R., barrier reef developed from fringing reef by subsidence to sea-level 2; A, the atoll which succeeds the barrier reef after volcano disappears below sea.

Many barrier reefs and atolls have probably originated in this way. But there are others which have had a different history. The wave-attack on some volcanic islands has produced broad wave-cut terraces which are extended seaward as wave-built terraces. The island of Tutuila, American Samoa, for example, is completely surrounded by such a wave-made platform which averages two miles in width. Along the outer margin of this shelf a barrier reef developed, while at the same time a fringing reef grew up along the shore and a broad shallow lagoon remained between them. This barrier reef, therefore, did not pass through an earlier fringing-reef stage. Similarly, an atoll may be produced by the growth of a reef around the edge of a truncated volcanic cone, or other wave-cut platform, without requiring subsidence of any consequence, and without passing through a barrier reef stage.

CHAPTER VII

LAKES

A lake is an inland body of standing water larger than a pond or pool. For the most part lakes are separated from the ocean and are commonly above its level, though some, such as the Caspian Sea and the Dead Sea, are below. Though called seas, the Caspian, Dead Sea, Sea of Galilee, Aral, etc., are truly lakes. The size which a body of water must have before it is called a lake is somewhat arbitrary, for in some places and by some people, a body of water is called a pond which elsewhere or by others would be called a lake.

Size and Depth. Most lakes are only a few square miles in area, though a few are very large, the Great Lakes aggregating about 95,000 square miles, and the Caspian approximating 170,000 square miles. Most lakes are shallow; small lakes whose depths are less than 50 feet vastly outnumber those of greater depth. Certain lakes, however, are known to be very deep — Crater Lake (2000 ft.), Tanganyika (2100 ft.), Nyassa (2300 ft.), Caspian Sea (3200 ft.), and Baikal (5618 ft.).

Origin of Lakes. Lake basins originate in many ways, some of which have been previously indicated. Most of them arise from processes of gradation.

1. *Rivers* meandering on their floodplains leave oxbow lakes. If a tributary brings a greater load to the main stream than the latter can carry, an obstruction is built which ponds the upstream water. Similarly a main stream may become so overloaded that it deposits across the mouth of a tributary, thus ponding the latter to make a lake. A river depositing material, either as an alluvial plain or a

delta, may enclose an area of the ocean and convert it into a lake basin, such as Lake Pontchartrain, on the delta of the Mississippi. Tulare Lake, in the central valley of California, is ponded by alluvial deposits from the mountains.

2. *Waves and currents* along a coast may deposit sand across the mouths of drowned valleys, or other bays, to produce lakes.

3. *Glaciers* excavate basins, or so deposit drift that basins are produced. Many lake basins in glaciated country owe their existence partly to glacial erosion and partly to deposition. Thus the basins of the Great Lakes were formed partly by digging out their lower parts and partly by building up their rims. A valley glacier sometimes obstructs the water of a tributary valley so that a lake is formed. Continental glaciers have not uncommonly ponded water between their outer margins and outlying highlands.

4. *Landslides* are known to have obstructed valleys to form lakes.

5. *Solution* of limestone gives rise to lake basins such as those in the limestone-sink areas of Tennessee and Florida.

6. *Weathering* of an area of rock more than the surroundings may so loosen the rock that it is blown away, leaving a depression.

7. *Wind* may scour a depression, or may pile up material, and construct a lake basin. Sand dunes in places have migrated into valleys, ponded the waters, and produced lakes.

In addition to those produced by these processes of gradation, lake basins are produced also by vulcanism and diastrophism. Lava flows obstruct valleys to produce lakes. During cooling, the surface of lava flows in some places sinks so as to produce basins which are later occupied by lakes. In the craters of inactive volcanoes lakes are not uncommonly found. Crater Lake, in Oregon (Fig. 249), about five miles in diameter and nearly two thousand feet deep, occupies the *caldera* believed to have been made by the sinking of the top of a volcanic mountain. This basin, while closely related to vulcanism, was actually the result of diastrophism. Other diastrophic movements, such as warping and folding, may produce depressions in which water accumulates to form lakes. Where rocks of the lithosphere are displaced along fractures, lake basins may also be formed (Fig. 251).

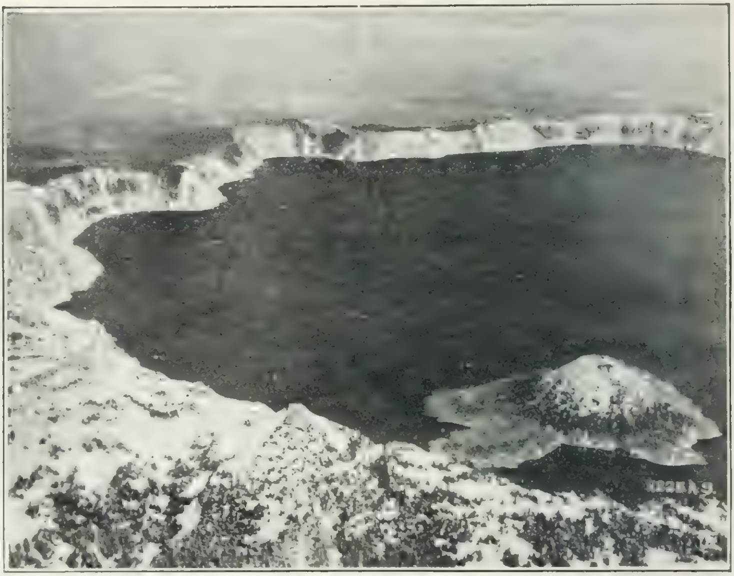


Fig. 249. — Crater Lake. (Courtesy U. S. Army Air Serv.)

Changes Taking Place in Lakes. The processes in operation in lakes are easily observed and readily understood.

Filling. Lake basins are constantly being filled. (1) The waves wear the shores, and the material thus derived is transported, assorted and deposited as in the sea, and all the topographic forms resulting from erosion or deposition along the sea-coast are reproduced on their appropriate scale in lakes. (2) Sediment-laden water of rivers and sheet-wash entering lakes must leave their burden of gravel, sand, and mud in the basin, for the outflowing water is clear. (3) Winds blow dust and sand from the land into lakes. (4) Animals of many sorts inhabit lakes, and help to fill up the basins with their bones and shells, which sink to the bottom. (5) Plants grow in lakes, particularly around shallow borders, and as they die their substance helps fill the lake basins. These processes tend to fill lake basins and so eventually to destroy lakes. But at the same time another process is actively engaged in destroying lakes.

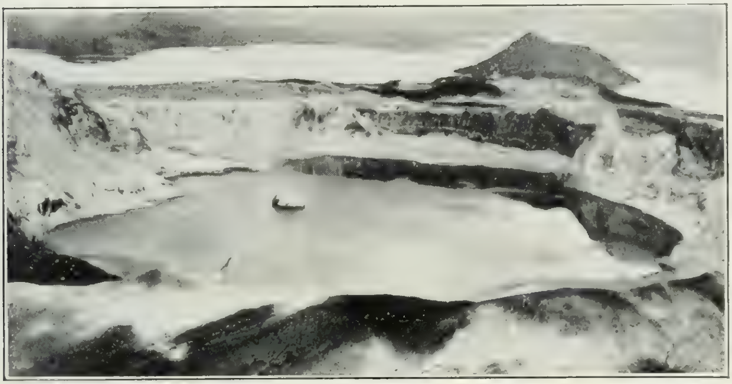


Fig. 250. — Crater of Taal Volcano Philippine Islands. (U. S. Army Air Serv.)

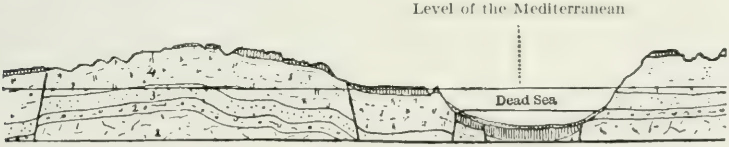


Fig. 251. — Cross-section of Palestine showing the relations of the Dead Sea. (After Blankenhorn, from Geikie's *Earth Sculpture.*)

Lowering of Their Outlets. The water flowing out of lakes which have outlets is constantly cutting down the levels of the outlets. This process tends to drain the lakes. Since, however, running water cannot cut below the level of base-level, only lakes whose bottoms are above this level can be entirely destroyed by the erosion of running water.

Fate of Lakes. Practically all lake basins are being filled all the time. Lakes above the level of base-level are being lowered by the lowering of their outlets. It thus becomes evident that all existing lakes will eventually be destroyed by one or the other of these processes or by a combination of the two. As geologic time is measured, lakes are ephemeral features. It does not follow, however, that lakes will cease to exist, for new ones will undoubtedly be formed to take the place of those destroyed.

Lacustrine Deposits. The beds of sediment deposited in lakes are similar in kind, in structure and in disposition, to beds of sediment laid down in the sea, though river-borne sediment is more commonly concentrated into deltas, since waves and shore-currents are less effective in lakes than in the sea. Finely stratified or laminated deposits of silt and clay are very common products of lake deposition (Fig. 252). Even the limestone made in the sea has its correlative in limestone made in some lakes. Some of it was made of the shells of fresh-water animals which throve where the inwash of terrigenous sediment was slight, some of it from the calcareous secretions of plants, and some of it was precipitated from solution. While still soft, such deposits are called *marl.*

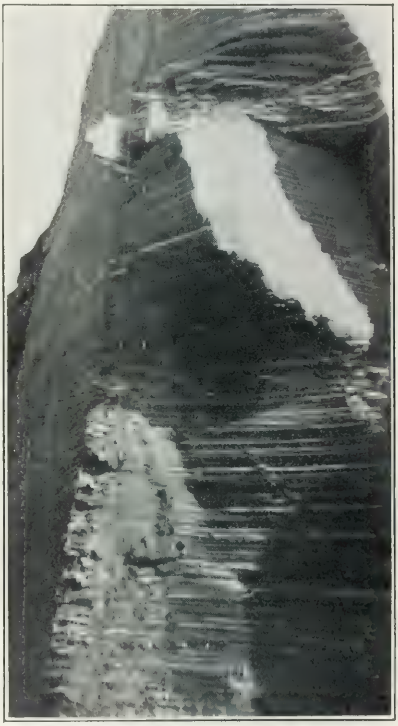


Fig. 252. — Laminated lake clays. Willimansett, Mass. (Margaret B. Fuller.)

Salt, gypsum, borax, and iron-ore deposits are also made in some lakes.

Extinct Lakes. The former existence of lakes where none now exist may be known in various ways. If a lake basin has been filled, its former area is a flat which bears evidence of its origin in its composition, its structure, and in places its fossils. Such a flat is commonly so situated topographically that the basin would be reproduced if the lacustrine deposits were removed. To this general rule there are exceptions, as where a glacier formed one side of the basin when it was filled. If, on the other hand, the lake was destroyed by the lowering of its outlet, or by the removal of some barrier such as glacier ice, or by desiccation, shore phenomena, such as beaches, terraces (Fig. 253), spits, deltas, etc., may be found even though there is no well-developed flat corresponding to the bed of the lake. In time, such features are destroyed by subaerial erosion, so that they are most distinct soon after a lake becomes extinct.

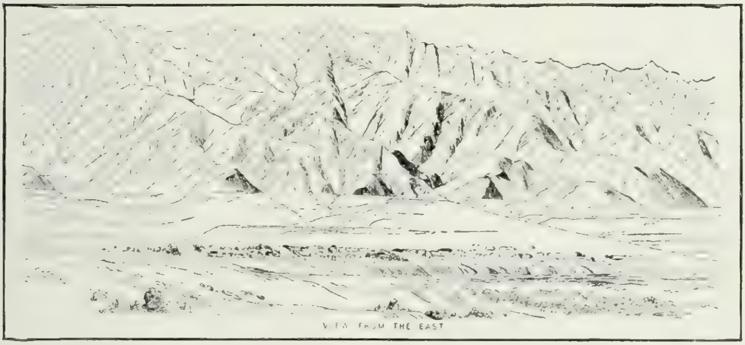


Fig. 253. Shore terraces of Lake Bonneville, Wellsville, Utah. (Thompson and Holmes.)

Salt Lakes. A few lakes, especially in arid or semi-arid regions, are salt, and others are "bitter." Besides common salt, salt lakes usually contain magnesium chloride, and magnesium and calcium sulphates, as well as some other mineral substances. "Bitter" lakes usually contain sodium carbonate, as well as sodium chloride and sulphate, and in some instances chiefly borax (Fig. 254). The degrees of saltiness and bitterness vary from freshness on the one hand to saturation on the other. The water of the Caspian Sea (lake) contains, on the average, less salt than that of the ocean (3.4%); that of Great Salt Lake contains about 18%; that of the Dead Sea, about 24%; and that of Lake Van, in eastern Turkey, the densest body of water known, about 33%.

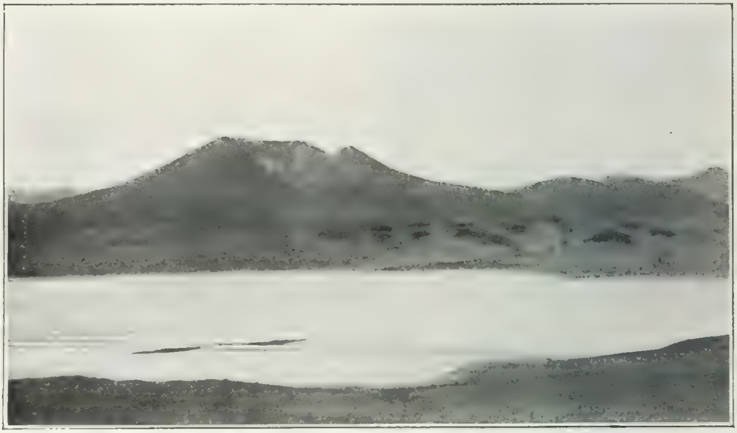


Fig. 254. — Borax lake of Ascotan, Chilean Andes. Borax derived from the crater in the background. (R. T. Chamberlin.)

Many salt lakes, such as the Dead Sea and Great Salt Lake, are descended from lakes which were fresh, while others, like the Caspian Sea, are probably isolated portions of the ocean. Lakes of the former class have usually become salt through a decrease in the humidity of the region where they occur, involving perhaps also changes in the outlets. The inflowing waters bring in small amounts of saline and alkaline matter, which is concentrated as evaporation takes place. The concentration may go on until saturation is reached or until chemical reactions cause precipitation.

Deposits of salt and other mineral matter once in solution are forming in some salt lakes at the present time, and considerable formations of the same sort have been so made in the past. Buried beneath sediments of other sorts, beds of common salt or of other precipitates are preserved for ages. Lime carbonate has been precipitated in quantity from some extinct lakes (Fig. 255).

Lakes which originate by the isolation of portions of the sea are salt at the outset. If inflow exceeds evaporation, they become less and less salty and, if there is an outlet, they may ultimately become fresh; otherwise they remain salt. If evaporation exceeds inflow they diminish in size and their waters become more and more salt or bitter. Any salt lake may become fresh by increased inflow of fresh water and the establishment of an adequate outlet which gradually carries away the salt.

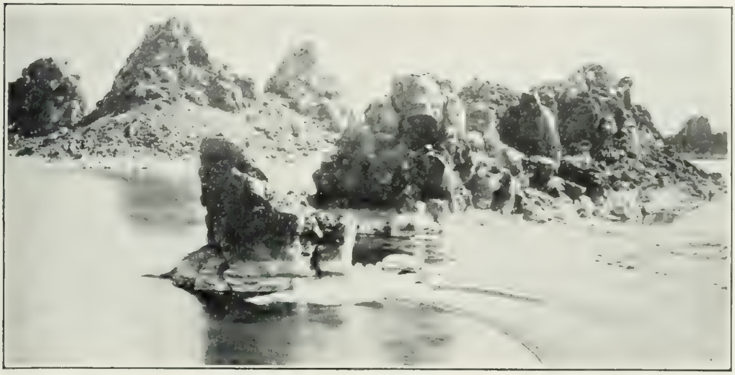


Fig. 255. — Tufa domes, Pyramid Lake, Nev. (Fairbanks.)

Indirect Effect of Lakes. Lakes tend to modify the climate of the region where they occur, both by increasing its humidity and by decreasing its range of temperature. They act as reservoirs for surface waters, and so tend to restrain floods and to promote regularity of stream-flow. They clear the waters which enter them by allowing their sediments to settle, and consequently influence the work and the life of the waters flowing from them.

CHAPTER VIII

DIASTROPHISM

It is evident from our study of the agents of degradation that land is being constantly lowered. Given time enough, the land of the globe would be brought below sea-level and so destroyed. The fact that land exists today means that another process is in operation by which the relief of the lithosphere is increased and land rejuvenated. Movement of one part of the lithosphere with respect to another is known as diastrophism. Its activity, both present and past, becomes apparent when many phenomena are scrutinized.

The evidences of diastrophic movements are strikingly shown along coasts, for sea-level is the datum plane from which land elevations are measured. Raised beaches and sea-caves, as well as drowned and submerged valleys, show that there has been movement of the land with respect to sea-level (Figs. 256 and 228).

On the other hand, it is evident that sea-level itself may not remain stationary, but may either rise or fall. The land would appear to rise if the sea-level should sink, while the land would appear to sink if the sea-level should rise. Filling in by sedimentation, the building up of the ocean floor by submarine volcanoes or lava flows, or movements of the sea bottom, may cause sea-level to change. Simply from the evidences along coasts, it is commonly possible to detect only *relative* movement of land and sea.

EVIDENCES OF RELATIVE MOVEMENT OF THE LAND

1. The ruins of the ancient temple of Jupiter Serapis stand on the coast near Naples, Italy. Judging from inscriptions the temple was above sea-level as late as 235 A.D. When

rediscovered in 1749, there were several columns still standing. Their lower parts were buried to a depth of 12 feet by marine sediment, while for an additional 9 feet the columns showed holes evidently bored by small marine animals. During the intervening time the site of the temple must have sunk relative to sea-level at least 21 feet, and then have risen again to bring it above the level of the Mediterranean.

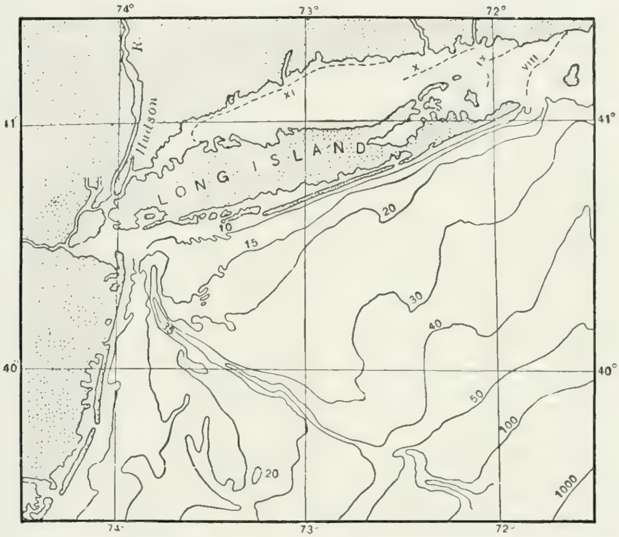


Fig. 256. — The submerged valley which has been interpreted as the continuation of the Hudson Valley. The position of the valley is indicated by contours. (Data from Coast and Geodetic Surv.)

2. Scandinavia is warping. Measurements in the northern part in places show a rise of about $2\frac{1}{2}$ feet per century. At Malinö, on the other hand, excavations reveal ancient paved streets eight feet below the level of the sea.

3. Relative rise of the land is clearly shown by the presence of beds of marine shells of living species high above the sea. Such beds are found in Sweden and in North Greenland, where they occur at altitudes of 100 to 200 feet. Darwin observed similar beds up to an elevation of 1300 feet on the west coast of South America. Many other cases of

similar kind are known from other parts of the world. Sedimentary rocks containing ancient marine fossils are to be found in some of the highest mountains of the earth.

4. On the other hand, land vegetation, as well as human structures, is to be seen in many places below the surface of the sea.

SHRINKAGE OF THE EARTH

In many places rocks which were originally flat are now seen to be folded, crumpled, and broken. Where folding or thrusting has taken place, the inference is that there has been shortening of the earth's circumference in that place. The shortening represented by the folds of the Alps has been estimated recently by Heim at from 125 to 175 miles. The shortening for the Appalachians, in Pennsylvania, not including the crystalline belt on the east, was 15 miles; including the belt of crystalline rocks east of Harrisburg it was probably more than 100 miles. The folding of the Rocky Mountains north of Denver, in Colorado, involved a surface shortening of about 8 miles, while farther north cross-sections show the shortening to have been much greater. Such estimates of circumferential shortening imply considerable radial shortening of the globe for each of these periods of folding. During the better-known geological eras there have been many such periods of folding; while in the very ancient rocks there are records of even greater movements resulting in greater folding. To produce this great amount of folding there apparently has been notable shrinkage of the earth.

Causes of Shrinkage. *Cooling.* It has long been held that the earth, once hot, is progressively cooling. If this were true this process would produce contraction. To test this it has been estimated from the conductivity of rock and the observed thermal gradients below the surface, that in 100,000,000 years the temperature would be lowered 10° C. (18° F.).[1] Another estimate gives the larger figure of 45° C. (81° F.).[2] For a loss of 10° C. the circumferential shortening would be 1.6 to 2.35 miles, while a cooling of 45° C. would give only 7.27 to 10.5 miles. These were quite inadequate to account for the shrinkage implied by the mountain folds and other deformations seen at the surface.

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[1] Tait, *Heat*, p. 225.

[2] Daniells, *Physics*, p. 407.

Furthermore, we cannot be certain that the interior of the earth has been growing colder during known geologic time. Heat is being developed within the earth in several ways, and it is possible that it is being produced as rapidly as it is conducted to the surface, or even more rapidly, in which case the interior of the earth would be getting hotter. It would be unsafe, in the present state of knowledge, to say that there has been any cooling of the earth since the time when the earliest rocks now accessible at the surface were formed. We must look for other causes of shrinkage.

Rearrangement under Pressure. According to current theory the interior of the earth is believed to be largely heterogeneous in constitution. It is considered to be made up of metallic and stony material together with gases and liquids admixed and dissolved. Acting upon this material, gravity produces tremendous pressures which attain something like 45,000,000 pounds per square inch at the center. Under these great pressures the material of the interior of the earth is subject to rearrangement and compacting. Ordinary physical compacting must be great, but in addition it is probable that molecular rearrangement, under the intense pressure and heat of the interior of the earth, produces new compounds. Further, it is a general law that re-aggregation under pressure takes the densest form available. A simple illustration of the decrease in volume by combination is seen in the case of water. Two volumes of hydrogen unite with one volume of oxygen to make *two* volumes of water. Hence re-aggregation in the interior and the building of denser molecules mean increased density and decreased volume. This may be constantly going on and constantly causing shrinkage in the interior.

The present theories of the constitution of matter are extremely interesting and important to the student of geology. Matter is made up of molecules. These are very small. It has been said that there are roughly as many molecules in a cubic centimeter of gas at normal pressure as there are grains in a cubic mile of sand. The numbers in a solid are greater still. Molecules are composed of atoms. For instance, as we have seen, two atoms of hydrogen are combined with one atom of oxygen to produce a molecule of water (H_2O). An atom itself is made up of a central nucleus, bearing positive electric charge, about which revolve one or more unit charges of negative electricity, known as *electrons*.

The size and space relations of atoms are very significant. While atoms are exceedingly small, they are, in comparison with the size of the nucleus and electrons which compose them, tremendously large. A familiar comparison might be given. If the nucleus of a hydrogen atom be likened to a small lead shot on one of the buildings in New York City, the electron should be likened to a fifteen-foot balloon revolving in an orbit as far out as Philadelphia.

When the "solid" earth is considered, it must be remembered that it is made up of these minute electrical charges tremendously deployed through space. This gives a suggestion of the vast theoretical compressibility of earth-matter, and should be kept in mind, though the extent to which this has played a part in the actual shrinkage of the earth cannot as yet be told. In the light of present knowledge, atoms of lower order, under the great compression in the earth, may be converted into those of higher order. The process is *endothermic* (using up heat) and the heat may be liberated under lesser pressure. It then follows that gravitational, molecular, atomic, and sub-atomic agencies have been, and still are, at work, tending to increase internal densities and so to produce contraction of the earth.

Rise of the Lava. Considerable portions of the surface of the earth are made up of igneous rock which rose from within the earth. The space once occupied by this lava must have been filled by the inward movement of the surrounding rocks. Notable distortions may have arisen in this way, but these are the indirect results of the operation of other forces.

The Effect of Heat. Concentration under the influence of pressure constitutes compression. Compression generates heat. Heat, in this case, is represented by the more rapid vibrations of the molecules resulting in the tendency of the material to expand and consequently to resist compressive movements. When this heat has been conducted to the surface and thence lost by radiation, or has been consumed by internal organization, compression may proceed further. Loss of heat is a secondary cause of contraction. According to this view, diastrophism results from the disturbance of the equilibrium maintained between contractive and resisting forces.

Periodicity. Concentration is apparently in progress all the time. But the solid earth is elastically rigid (i.e., the molecules hold their places as in solids — but yield elastically), resisting movements which would otherwise be manifested in continuous, slow warpings. It is not until the *stresses*, produced in concentration, accumulate in sufficient magnitude to overcome the strength of the outer part of the earth, that movement sets in, *strains* are produced, and diastrophism becomes evident. Periods of great disastrophism are used to demark the eras of geologic history.

I. RADIAL MOVEMENTS (EPEIROGENIC)

Condensation and resulting contraction of the earth produce movements toward the center. Since the direction of these movements is, in general, parallel with the radius, they may well be called *radial movements*.

The surface configuration of the lithosphere leads to the conclusion that radial movements have taken place at different times and to varying amounts in different parts of the earth. Thus the ocean-basin segments differ conspicuously from the continental segments; and plateau segments differ from lowland segments. One of the leading views is that the solid earth is made up of wedge-shaped segments pointing inward and that its dynamics involves the relative movements of these segments.

Wedge Dynamics. Measurements of the pull of gravity reveal the fact that the ocean segments are of greater specific gravity than the continental segments. Additional evidence of the greater oceanic density is furnished by seismic phenomena and by the average density of the accessible rocks. Being denser and larger, for they comprise about 68% of the surface of the globe, the ocean segments are the master segments.

Densities of Continents and Ocean Floors[1]

(Water-free Basis)

Area	*Height*	*Density*
North America	700 meters	2.782
South America	580 "	2.783
Europe	300 "	2.790
Africa	650 "	2.780
Asia	950 "	2.769
Australia	350 "	2.814
Atlantic	–4,116 "	2.888
Pacific	–4,520 "	3.087

When, in the process of earth condensation, the stresses have grown sufficiently so that the material of the outer portion of the earth can no longer withstand them, yielding begins. This yielding consists in the inward movements of the wedges, and the ocean segments, because heavier, take the lead in descending. The continental areas between them, being lighter, sink less in the movement toward the center.

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[1] H. S. Washington, "Isostasy and Rock Density," *Geol. Soc. Amer. Bull. 33* (1922), p. 404.

As the ocean segments sink most, the ocean basins become deeper and increase their capacity. This results in drawing the shallow seas off the continents and increasing the land area. As the ocean segments sink most, they crowd in ahead of, and under, the lighter and smaller continental segments, producing the effect of apparent raising of the land masses. The continental segments are squeezed upward relative to the oceanic segments, though probably not with respect to the center of the earth (Fig. 257). This, in short, is the *continent-making movement*, by which eroded continents are restored.

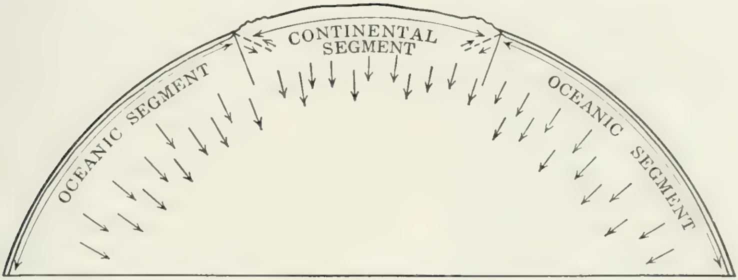

Fig. 257. — Diagram to illustrate deformation of a continent. Arrows pointing toward center of earth indicate general condensation. Long arrows represent tangential compressive stress. All segments sink, but the continent is wedged up somewhat relative to the oceanic segments. Dash lines represent shear under the continents. Vertical lines represent relations of segments before diastrophism.

II. CIRCUMFERENTIAL MOVEMENTS (OROGENIC)

As has been shown, diastrophism is expressed primarily by movements toward the center of the earth, i.e., radial. But as the segments sink into smaller circumferences there is more and more crowding. Lateral compression, with horizontal thrusting, develops throughout the outer portion of the globe within both continental and oceanic segments (Fig. 257). But as the oceanic segments are most active in the sinking, the effects of the crowding are most felt along the borders of the continental masses.

The horizontal thrusting causes wrinkling into mountain ranges in those belts where the intensity of the thrusting forces becomes too great for the strength of the rocks to resist. Two general conditions determine where the horizontal compression becomes too great for the resistance of the rocks. They are: (1) places where the compressive forces are exceptionally intense, and (2) regions of weaker rocks, particularly in strips or troughs where exceptional thicknesses of soft sediments have accumulated, and where the inclined attitude of the rock layers makes them less resistant.

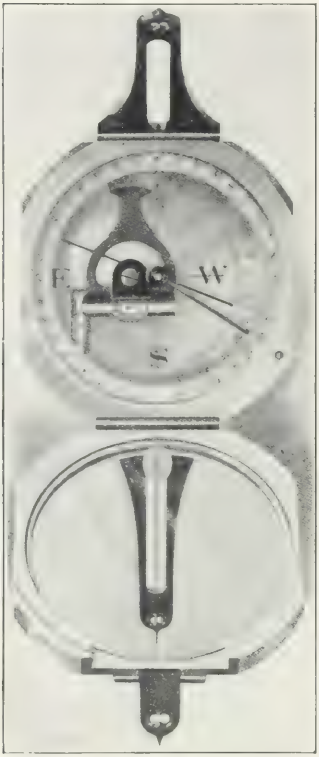
Fig. 258. — Brunton pocket transit and clinometer.

Soft sediments, on the average, accumulate in greatest thickness near the margins of the continents, for it is here that rivers, wind, and waves make their greatest deposits. While thrusting forces develop throughout the entire continental masses, they are naturally most intense near the *borders* where, in addition to the general thrusting, there is the further crowding by the oceanic segments. This *extra thrust* from the master segments is the controlling or "trigger" effect which causes the wrinkling most commonly to occur near the margins of the continents. Hence we find that most of the mountain ranges have formed along the borders of the continents and more or less parallel to these borders.

If the *yielding* has been concentrated in a comparatively thin shell which has been intensely deformed, a *mountain-system* results. If the yielding has been distributed through a much thicker shell, which perforce has been less intensely deformed, the region is bulged up into a plateau-like area. One of the chief processes in mountain-building is folding.

Folding

The original attitude of beds, whether formed by water or by lava flows, commonly departs but little from horizontality. Both kinds of deposits, however, have been formed in places on considerable slopes. Modifications of the original attitude result from earth movements, and the measurement of these modifications is an important feature of geologic field study. The amount that a bed is tilted from the horizontal is called the *dip*, and is measured in degrees by means of an instrument known as a clinometer. In measuring the dip, the greatest angle of inclination is always taken. The bearing of a line at right angles to the dip, i.e., a horizontal line along the outcropping edge of a dipping bed (Fig. 259), is called the *strike*, and is recorded by compass readings (e.g., N. 40° E.). When beds incline in a single direction, they form a *monocline*. When they are arched up, as in an upfold, they form an *anticline*. There are anticlines in which the beds have only very gentle dips; those in which the beds dip steeply (Fig. 260); those in which the beds, or *limbs* of the fold, are pressed tightly to-

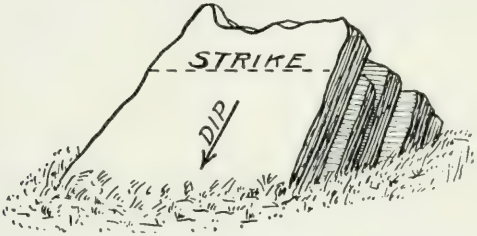

Fig. 259. — Diagram illustrating dip and strike. (Jas. Geikie.)

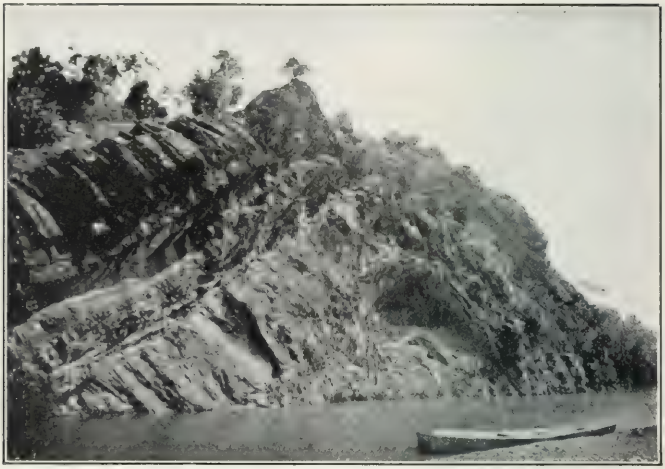

Fig. 260. — Anticlinal fold, near Hancock, Md. (U. S. Geol. Surv.)

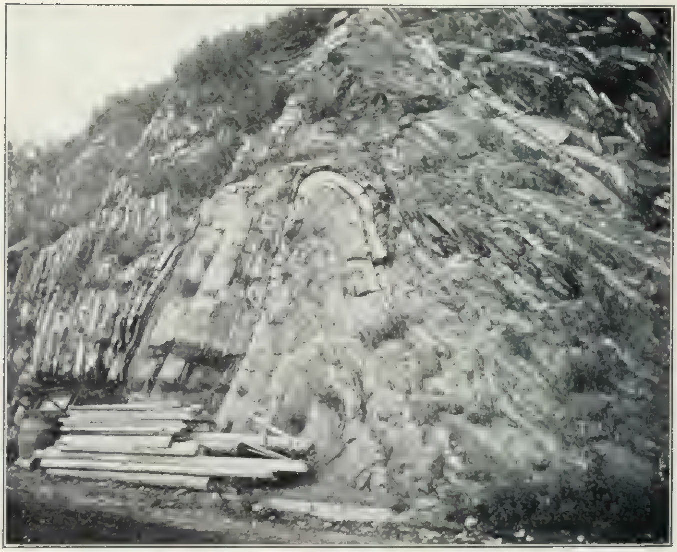

Fig. 261. — Anticlinal fold, near Levis Station, Quebec. Limbs pressed tightly together. (U. S. Geol. Surv.)

gether (Fig. 261); and those in which the thrusting has been so great that the fold is *overturned* or *recumbent* (Fig. 262). If the axis of the fold plunges, the structure is known as a *plunging anticline*, while if the axis plunges at both ends, a *doubly plunging anticline* is formed. If the rocks dip away in all directions (quaquaversal dip) the structure is known as a *dome*. In oil-field usage a *closed anticline* is a dome, or a doubly-plunging anticline, under which oil may accumulate, while an *open anticline* is one whose axis is either horizontal or plunges in but a single direction. The downfold corresponding to an anticline constitutes a *syncline* (Fig. 263). A syncline, like an anticline, may be gentle or closely pressed; upright or inclined; horizontal or plunging.

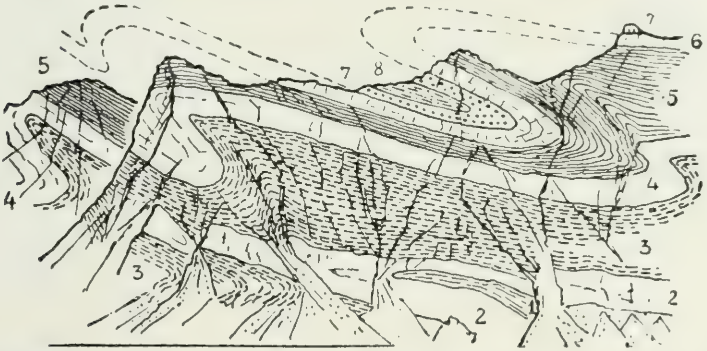

Fig. 262. — Overturned anticlines. South side of Diablerets, Switzerland. (After Renevier in Heim. Geologie der Schweiz.)

When strata have been strongly folded, so that the limbs of the folds are essentially parallel to each other, the folds are *isoclinal* (Fig. 264).

When a considerable tract is bent so as to form a great broad arch, the term *geanticline* is sometimes applied to it. Similarly a broad trough of large extent, formed by downwarping, particularly one in which sedimentary deposits accumulate to exceptional thickness, is called a *geosyncline*. A composite structure, consisting of several or many smaller anticlines and synclines, is called an *anticlinorium* (Fig. 265) if the structure as a whole is an upfold, or a *synclinorium* (Fig. 266) if the whole structure is a downfold. The use of these terms may be illustrated from the geologic history of the eastern United States. The Appalachian geosyncline, a downwarped trough filled with sediments in early geologic times, was later folded into the Appalachian anticlinorium, giving rise to the Appalachian Mountains.

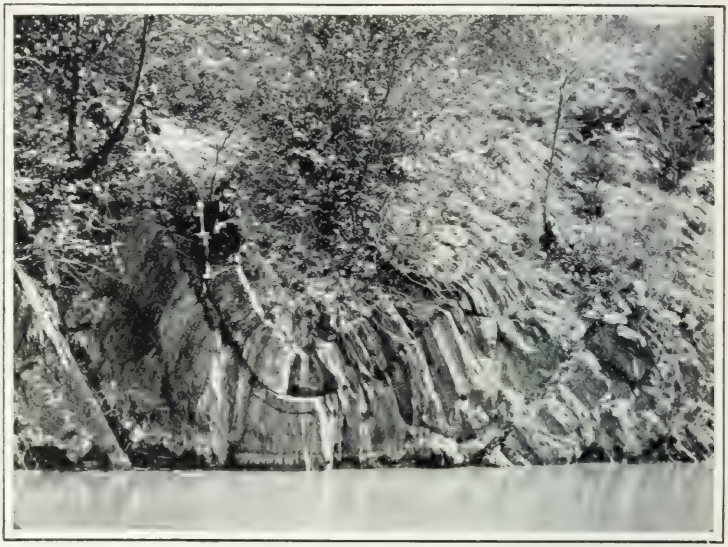

Fig. 263. — Syncline, C. and O. Canal, 3 miles west of Hancock, Md. The beds are shale and sandstone. (Walcott, U. S. Geol. Surv.)

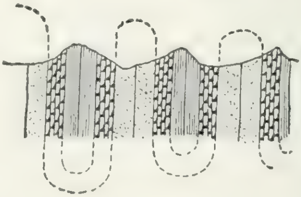

Fig. 264. — Isoclinal folds. (Van Hise, U. S. Geol. Surv.)

As found in the field, folds are usually much eroded, and some are completely truncated. The structure is then determined by a careful observation and record of dips and strikes. On the field map, the record may be made as shown in Figures 267 and 269. Here the arrows point in the direction of the dip, while the other lines represent the direction of strike. Applying this method, the structure shown in Figure 267 represents a syncline, and that in Figure 269 an anticline. In cross-section, the structure represented by Figure 267 would appear as in Figure 268. Figure 271 shows a doubly-plunging anticline. Figure 272 shows a combination of synclines and anticlines; and Figure 273 shows a cross-section along the line *ab* of Figure 272. The outcrops of rock where dip and strike can be determined may be few and far between, but where they are sufficiently near one

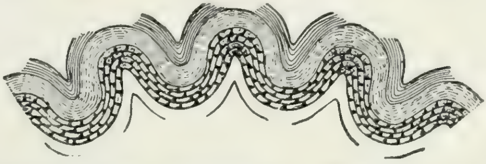

Fig. 265. — Anticlinorium, diagrammatic. (Van Hise, U. S. Geol. Surv.)

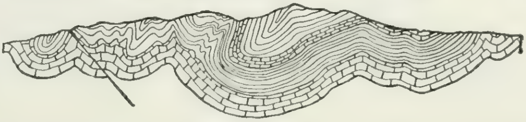

Fig. 266. — Synclinorium, Mt. Greylock, Mass. (Dale, U. S. Geol. Surv.)

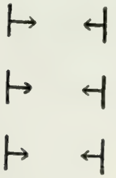

Fig. 267

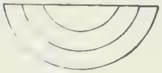

Fig. 268

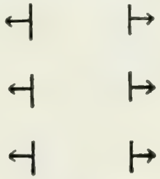

Fig. 269

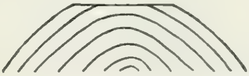

Fig. 270

Fig. 267. — Map record of dip and strike, showing synclinal structure.
Fig. 268. — Diagram illustrating the structure corresponding with Fig. 267.
Fig. 269. — Map record of dip and strike showing anticlinal structure.
Fig. 270. — The structure of the area shown in Fig. 269 in cross-section.

another, the structure of the region, as shown in Figure 272, may be worked out even though the land surface is flat.

Much the larger portion of the earth's surface is occupied by beds that depart little from their original horizontal attitude, but in mountainous regions the beds have usually suffered, in the course of the deformations that gave rise to the mountains, various degrees of bending, folding, crumpling, and crushing. In places the yielding of the rocks has been so great that not only their structure, but

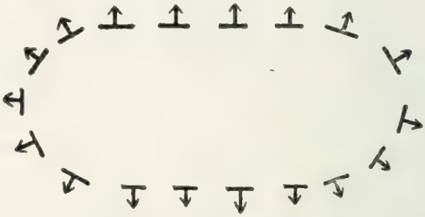

Fig. 271. — Map record of dip and strike showing plunging (dipping down at ends) anticline.

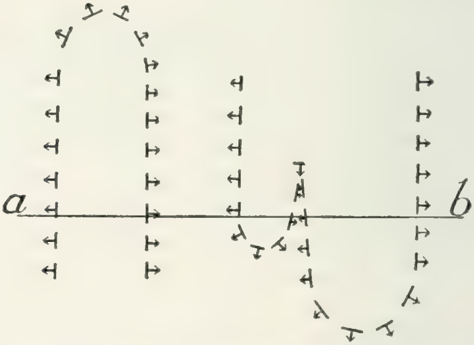

Fig. 272. — Map record of dip and strike showing complex structure.

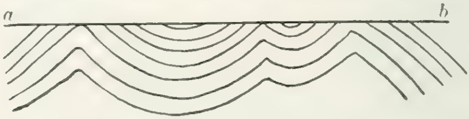

Fig. 273. — Cross-section of Fig. 272, along the line *ab*.

their internal texture, and even their mineral composition have been altered. Such alteration is known as *metamorphism* (dynamo-metamorphism).

Metamorphism

When rocks made up of clastic particles are compressed in one direction, and are relatively free to expand in others, there is a tendency for particles that are already elongated to turn so that their longer axes are in the direction of easiest relief, and for particles in general to become more or less flattened, commonly (though not necessarily) at right angles to the direction of compression (Fig. 274). But much more important than this is the development of new crystals and new minerals within the rocks. The new minerals which commonly develop are such as form flattish or tabular crystals, particularly mica. A shale which contains no mica at

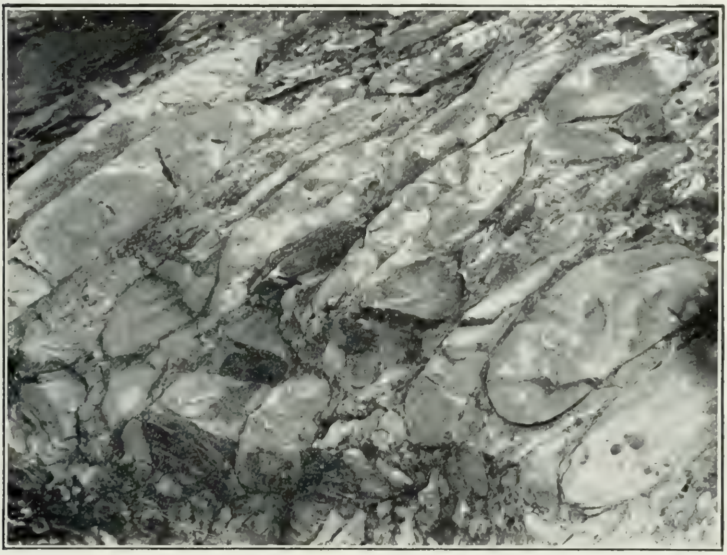

Fig. 274. — Figure showing the elongation of pebbles resulting from pressure. Carboniferous formation, Bancroft Place, Newport, R. I. (Walcott, U. S. Geol. Surv.)

the outset may undergo such extensive recrystallization as to develop 50 per cent of mica under extreme compression. This secondary mica, developed under the conditions of compression, will be so arranged that the longest dimensions of its crystals will be in the direction of easiest relief from compression and the shortest dimensions at right angles to the direction of relief. In other words, the flat flakes of mica have a nearly parallel arrangement.

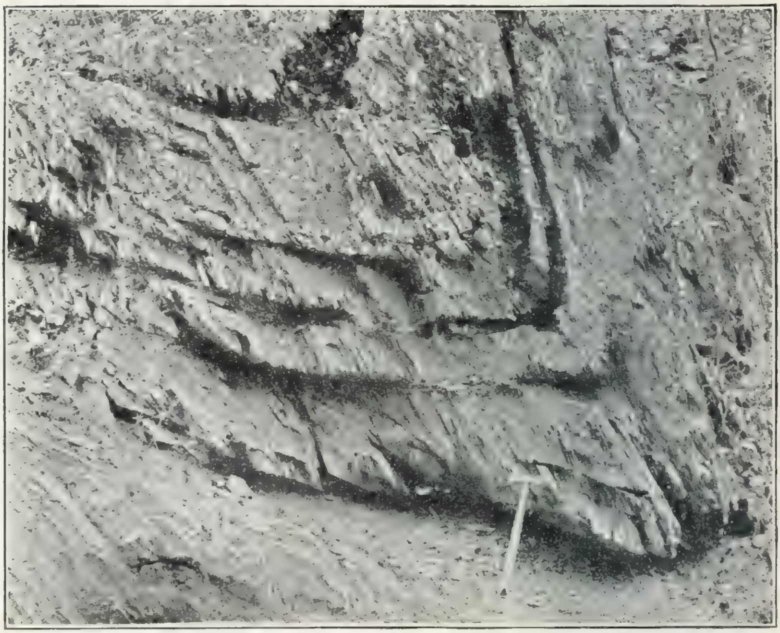

Fig. 275. — Slaty structure and its relation to bedding planes. Two miles south of Walland, Tenn. (Keith, U. S. Geol. Surv.)

As a result of this parallel arrangement of their particles, rocks so affected split more readily between the elongate and the flattish particles, than across them. In other words, the rocks *cleave* along planes parallel to the elongation of the rock mass, and break with difficulty, and with rough fracture, across the planes of cleavage. The condition thus induced is known as *slaty cleavage* (Fig. 8), and is best illustrated by roofing slate, which was originally a mud, later a shale, and finally assumed the slaty condition under strong compression. In a great many cases the original bedding may still be seen running across the cleavage planes developed by pressure (Fig. 275). Limestones, sandstones, and conglomerates do not develop platy minerals so readily as mudstones, and they usually take on only an imperfect cleavage.

In a slate the mica particles are minute and the cleavage surfaces are essentially planes. If, however, the mica crystals are larger, so as to be very evident to the eye, or if the cleavage surfaces are wavy, or both things are true, the rock is called a *schist* (Fig. 276) and the capacity to split along such surfaces is called *schistosity* or *foliation*. Schists are of many kinds, depending upon the proportions and kinds of minerals (mica, hornblende, etc.)[1] which are present. They may be produced by the metamorphism of either igneous or sedimentary rocks. Technically, schists contain little of the mineral feldspar. If feldspar is an important constituent and the rock is banded or foliated, it is called a *gneiss* (Fig. 278). Gneisses range from those with well-

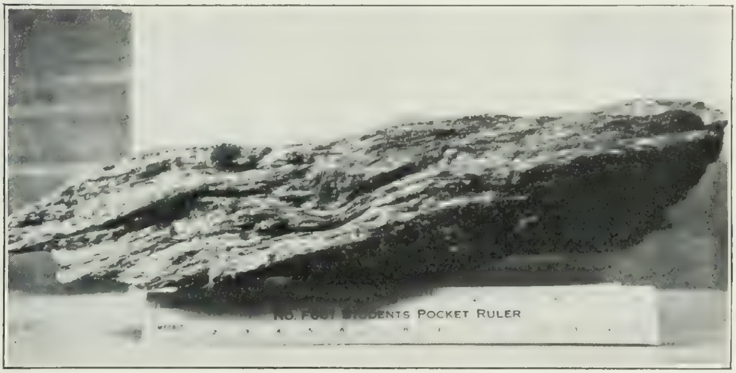

Fig. 276. — Schist.

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[1] Minerals will be discussed in the next chapter. For description see table beginning on p. 362.

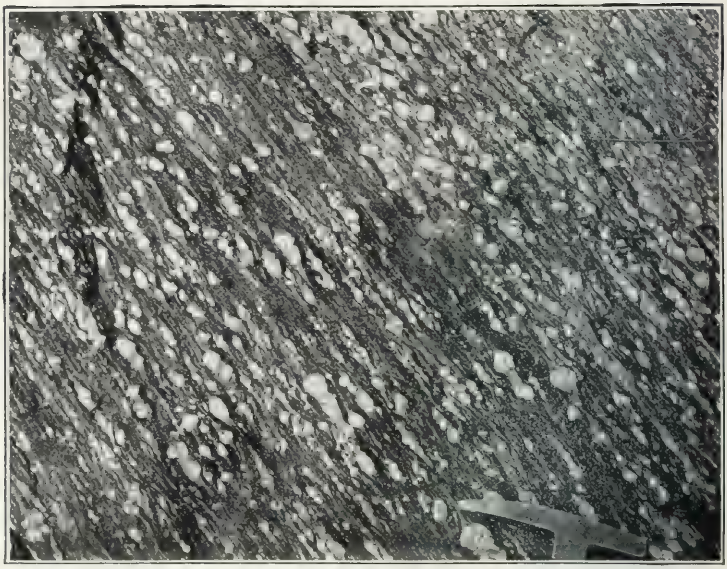


Fig. 277. — Porphyry rendered schistose by pressure. Near Green Park, Caldwell Co., N. C. (Keith, U. S. Geol. Surv.)

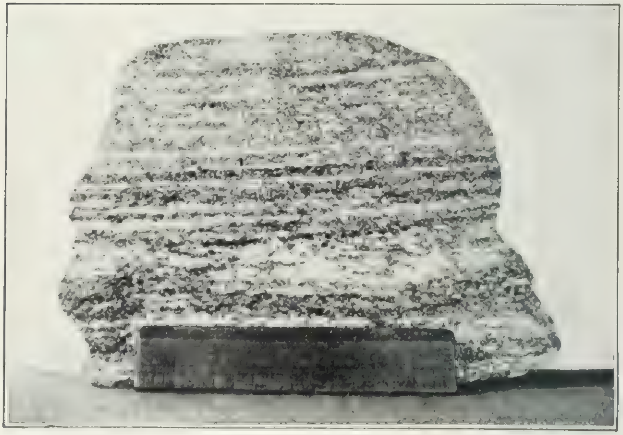


Fig. 278. — Gneiss.

developed schistosity (much mica) to those having only a banded texture due to the segregation of the minerals (principally quartz and feldspar) in bands. Gneisses are formed in several different ways. Some are metamorphosed granite rocks; some are metamorphosed sediments; while in others the banding represents lines of flow in a solidifying granite.

The alteration of rocks into slates, schists and gneisses under strong compression is facilitated by heat and the presence of water, which aid in the process of recrystallization. This is especially a chemical process. Under other conditions, intense pressure in a given direction may break down or crush the rock mechanically, producing an *autoclastic* (self-fragmented) rock.

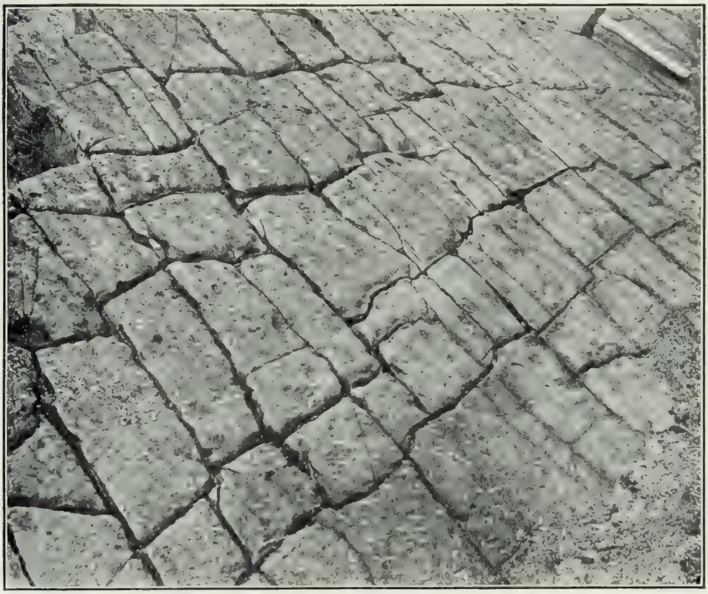


Fig. 279. — A surface of sandstone marked by numerous joints, chiefly in two rectangular sets. Near Kinghorn, Fife. (H. M. Geol. Surv.)

Joints. The surface rocks of the earth are almost universally traversed by deep cracks, called *joints* (Figs. 6 and 279).

In most regions there are at least two sets of joints, the crevices of each set being roughly parallel to one another. While the different sets may intersect at almost any angle, if there are only two sets, they intersect most commonly approximately at right angles. In regions of great disturbance, the number of sets of joints is often three, or even more. The joints of each set may be many yards apart, or but a few inches, or even a fraction of one inch apart.

In only slightly disturbed rocks the joints as a rule approach verticality, but in regions where the rocks have been notably deformed, the joint planes may have any position. In igneous and metamorphic rocks they may simulate bedding planes (Fig. 280). Some joints end at the plane of contact of two sorts of rock. Thus a joint extending down through limestone may end where shale is reached. Joints are frequently offset at the contact of layers or formations, and a single joint sometimes gives place to many smaller ones. All these phenomena are determined by the nature of the causal stresses and by the varying elasticity of various sorts of rock. Generally speaking, brittle, rigid rock is more

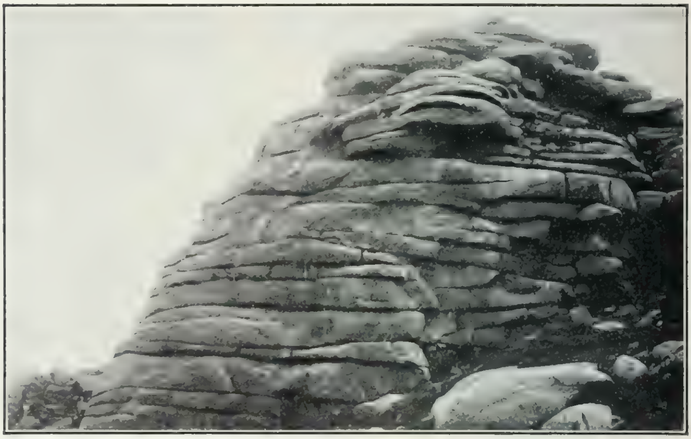


Fig. 280. — Tabular joints in granite. Summit of Goatfell Arran.
(H. M. Geol. Surv.)

readily jointed than that which is more yielding. Though some joints probably extend to greater depths than rock has ever been penetrated by man, they are believed to be relatively superficial phenomena even in the stronger rocks where they go deepest.

Joints may remain closed or they may gap. They may be widened by solution, weathering, etc., and may be filled, partly or wholly, by detritus from above, or by mineral matter deposited from solution (*veins*). Many ore veins are developed along joint planes. Exceptionally, open joints are filled by the intrusion of sedimentary material from beneath. This is the origin of the remarkable *sandstone dikes* of the West, the most striking of which are those of California. One of these dikes is nine miles in length. The sand of these dikes was forced up from beneath either by earthquake movements or by hydrostatic pressure.

Joints are produced either by tension (pulling apart) or by compression. Torsion (twisting) is also to be listed, though, in this process, tension is the important operating cause of the cracking. In the formation of an anticline tension joints will develop along the crest of the fold if tension goes beyond the limit of elasticity of the rock involved. If the axis of a fold is not horizontal, i. e., if it plunges, as it commonly does, a second set of tension joints may be developed roughly perpendicular to the first. A dome-shaped uplift, due to injection of lava from below, will show tension joints radiating from the center, as well as joints concentric with the periphery.

Shrinkage of the rock material is the cause of certain types of tension jointing. The cracks developed in mud where it has dried, and in shales and other sedimentary rocks upon consolidation, are examples. Most joints in regions of sedimentary rocks where there has been little deformation are probably tension joints.

Compression joints are found on the limbs of folds, adjacent to intrusions where there has been thrusting, and particularly in regions where other phenomena indicate regional compression.

Faults. If the rocks on one side of a joint plane, or fissure, are displaced relative to those on the opposite side, a *fault* is produced. The fracture surface along which the movement takes place, known as the *fault plane*, may be either horizontal, vertical, or in any intermediate position. The angle which the fault plane makes with the horizontal, as in the case of inclined strata, is the *dip*. In Figure 281 the distance *ab* is the *displacement*. The vertical component of the movement *ac* is the *throw*, while the horizontal component (*bc*) is the *heave*. If there is vertical displacement, the *upthrow* side will make a cliff or *fault scarp*. The exposed position of the scarp renders it more subject to erosion, and small ones soon disappear, but some scarps even of mountainous heights may be seen, as, for example, the east face of the Sierra Nevada, and the steep sides of many of the basin ranges of Utah and Nevada.

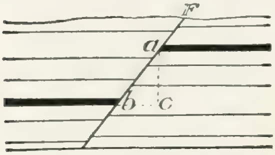


Fig. 281. — Diagram of a high-angle (normal) fault.

The rock on either side of a fault plane is commonly smoothed as the result of friction during movement. Such surfaces are *slickensides*. A slickensided surface has some resemblance to a glaciated surface, but is generally more glazed. The rock along some fault planes is crushed, making a *crushed zone* a few inches or a few feet wide. Such a zone becomes a passageway for ground-water, and so the site of ground-water work. Not a few such zones are "mineralized" and have become the sites of important mines.

Materials under tension commonly break essentially at right angles to the pull, while under compression they commonly break at angles more gently inclined to the pressure. In accordance with this principle of genesis, fractures are of two sorts — those formed under tension being at high angles to the effective force, and those formed by compression being at low angles to the effective force.

Faults involving vertical movement are therefore of two classes, *high-angle* (normal or gravity) and *low-angle* (reverse or thrust).

While some high-angle faults are produced by thrusting and some low-angle faults are the result of gravity, these are in a sense exceptional cases, as where injected lava faults the overlying rock. And it is because no present classification of faults is without serious objections, that the division into high-angle and low-angle faults is made. In the high-angle fault (Fig. 281) the fault plane makes a large angle with the horizontal, i. e., it is steeply dipping, and the hanging wall (overhanging side), in most cases, is the downthrow side. High-angle faults, as a rule, indicate an extension of strata, this being necessary to permit the dissevered blocks to settle to their new position. In the low-angle fault, on the other hand, the overhanging beds, in the majority of cases, have moved (relatively) up the slope of the fault plane, as though the displacement took place under lateral pressure. This is clearly shown where an overfold passes into a low-angle fault (Fig. 282).

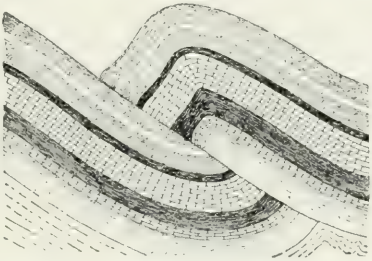


Fig. 282. — Diagram of a low-angle (reverse or thrust) fault. (U. S. Geol. Surv.)

In some low-angle faults displacement is very great. The eastern face of the Rocky Mountains near the boundary-line between the United States and Canada has been pushed over the strata of the bordering plains to a distance of at least 15 miles. Overthrusts of similar and even greater magnitude have been described as found in the Scottish Highlands, the Alps, the southern Appalachians and elsewhere.

Some faults branch into lesser faults. Others, instead of being a single fracture, are made up of a series of parallel planes of slippage near to each other, making a zone of faulting. Both are *distributive* faults (Fig. 283). Faults, when traced horizontally, are found to die out, some by passing into folds (Fig. 284), others by becoming distributive, and others by gradual lessening of the displacement due to the distributed compacting or stretching of the rock materials. At depth they probably die out in similar ways. A fault, if of more than a few feet in displacement, is probably the sum of numerous slight slippings extending over long periods of time.

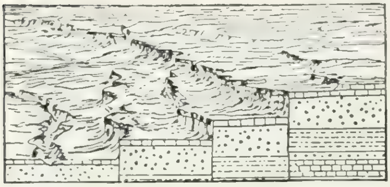


Fig. 283. — A branching fault. (Powell, U. S. Geol. Surv.)

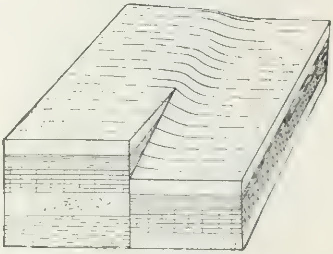


Fig. 284. — Diagram of a fault passing into a monoclinal fold.

Significance of High- and Low-angle Faulting

Faults afford an indication of the conditions of stress to which a region has been subjected, but caution must be exercised in their interpretation. High-angle faulting, as a rule, implies tension, while low-angle faulting implies thrust. But there may be high-angle faulting in regions of compressed and folded strata by subsequent stretching. And low-angle faults are found in regions of tension, but such cases are exceptional and usually local, as in the case of a large landslide where the angle of the slide plane becomes low toward the base of the slope.

In view of the current opinion that the outer part of the earth has been subjected to great lateral thrust as a result of contraction, it is well to make especial note of the fact that the faults which imply stretching are more abundant than those which imply compression. The prevalence of stretching is also seen in the abundance of gaping crevices, and of such veins as were open until filled by deposition. All these phenomena indicate that the surfaces of the continents are spreading and settling from the elevated positions produced by the major diastrophic movements. Times of settling and spreading naturally follow times of uplift, and tension faults may develop in places where at an earlier time thrust faulting prevailed. But, on account of the advantage of the low angles and the greater average displacement of thrust faults, it is probable that the crustal shortening shown by thrust faulting exceeds the crustal elongation shown by tension faulting.

Effect of Faulting on Outcrops

Faulting may bring about numerous complications in the outcrops of rock formations. In a series of formations having a monoclinal structure (Fig. 285), many changes may be introduced. Let it be supposed in the following cases that, after faulting, the surface has been eroded to planeness. It is to be noted that erosion shifts the outcrops down the dip of the strata. If the fault plane is parallel to the strike of the beds (*ab*, Fig. 285), and hence a *strike fault*, the outcrop of a given layer (*H*) may be duplicated (Fig. 286) or it may be eliminated altogether (Fig. 287). If the fault plane is parallel to the direction of dip (*cd*, Fig. 285), a *dip-fault*, the layer *H* will outcrop as in Figure 288, if the downthrow was on the far side. The outcrops of *H* are *offset*, the amount of the off-

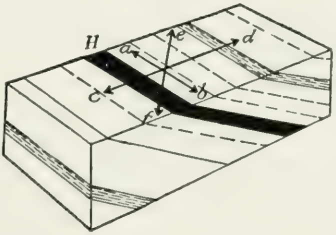


Fig. 285. — Diagram of an area of rocks with monoclinal structure. One layer notably unlike the others.

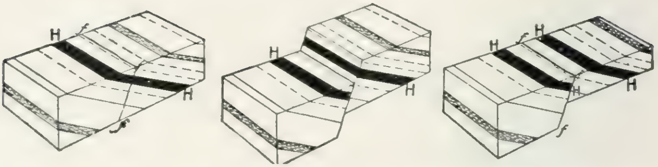


Fig. 286. — Same as Fig. 285 after (1) displacement by a strike fault, and (2) base-leveling. The outcrops of certain beds are repeated.

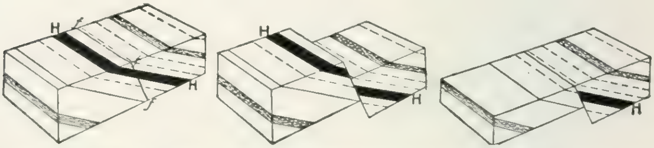


Fig. 287. — Diagram illustrating how a strike fault in such a structure as that shown in Fig. 285 may cause the outcrop of certain beds to disappear.

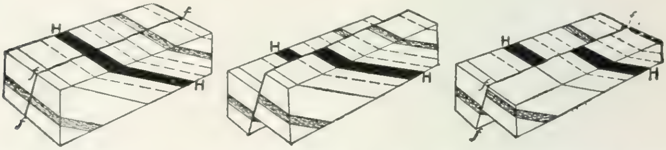


Fig. 288. — Diagram illustrating how a dip fault in the structure shown in Fig. 285 affects the outcrop when the downthrow was on the farther side of the fault plane.

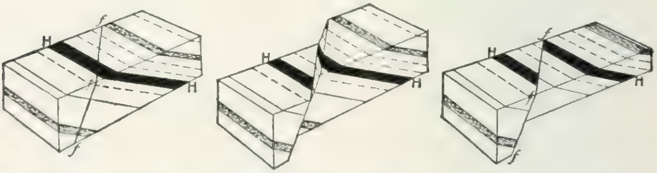


Fig. 289. — Oblique fault in the structure shown in Fig. 285. The downthrow was on the left side. The outcrop of layer *H* is offset with *overlap*.

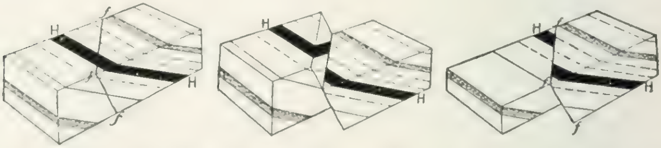


Fig. 290. — Same as Fig. 289, except that the downthrow was on the right side, and the offset is with a *gap* instead of an overlap.

set decreasing with increasing angle of dip, and increasing with increasing throw of the fault. If the fault is oblique to the direction of dip and strike (*ef*, Fig. 285), an *oblique fault*, the outcrop of such a layer as *H* will have the relations shown in Figure 289 if the downthrow was to the left, and those shown in Figure 290 if the downthrow was to the right. In the former case, it is said that there is *offset with overlap;* in the latter, *offset with gap*. The amount of the overlap and gap, respectively, increases with the increase of throw, and decreases with increase of dip. In all cases the outcrop (after degradation of the upthrow side) is shifted down dip, and shifted more on the upthrow side, where erosion is greater.

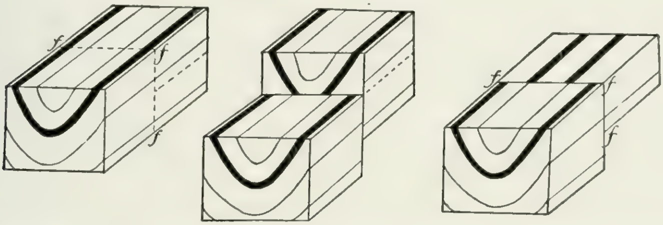


Fig. 291. — Diagram showing effect of faulting on the outcrops of synclinal beds.

If a fault crosses folds at right angles to their axes, the effect is to change the distance between the outcrops of a given bed on opposite sides of the fault, after the truncation of the folded beds. The distance is decreased on the upthrow side of a syncline (Fig. 291), and increased on the upthrow side of an anticline (Fig. 292).

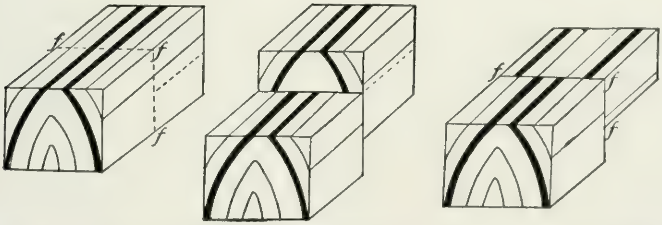


Fig. 292. — Diagram showing effect of faulting on outcrops of anticlinal beds.

Various other complications arise under other circumstances. Since fault scarps are not common, the detection and measurement of faults are usually based on the study of the relations of the beds involved, as just illustrated.

Faults of Horizontal Displacement

In the interest of clearness and simplicity the preceding discussion has concerned faults in which the direction of slip has been parallel to the dip of the fault plane, but faulting is not limited to such displacements. Horizontal displacements may take place parallel to the strike of a fault plane, with no vertical displacement; or there may be any combination of vertical and horizontal displacement. The 1906 California earthquake was produced by a slip along a vertical fault plane in which the horizontal movement reached a maximum of 21 feet, while the vertical component of the movement amounted to only about one foot. In some mines, where the walls of the shafts and tunnels afford excellent opportunity for determinations, horizontal movement is much more in evidence than vertical. Faults should, therefore, properly be considered in three dimensions. There are various displacements of rock bodies not mentioned above which are akin to faulting, if not to be regarded as such. For example, when strata are folded there is some slipping of layer on layer. Such movement between beds has been commonly referred to as *shear*.

III. ACCOMPANYING VIBRATIONS, EARTHQUAKES

The lithosphere is in a state of perpetual tremor. For the most part, these tremors are too slight to be sensible, though detectable by delicate instruments (seismographs). Some of them arise from disturbances within the earth-body, but many of them spring from the ordinary incidents of the surface, such as winds, tides, the rise and fall of atmospheric pressure, the rumble of traffic, the blasts of mines, and other causes. These incidental movements, while serviceable for demonstrating the elastic nature of the lithosphere, are not considered in this discussion.

Earthquakes

When tremors of appreciable violence spring from sources within the earth, they are classed as *earthquakes*. The causes are various; the most prevalent is probably the fracture of rocks and the slipping of rock masses on each other in the process of faulting. To the same class belong movements due to slumping, which is, in reality, superficial faulting. Tremors often attend volcanic eruptions. Such tremors are attributable to the sudden fracture and displacement of rock by the movements of lava, or to the operation of gas-pressure in various ways. Vibrations of a very minor sort also spring sometimes from the collapse of the roofs of caverns.

Depth of Origin: Focus. It is computed by seismologists that the average depth of origin of the large earthquakes is in the neighborhood of 25 miles. But some of the deeper foci have been variously computed to be at a depth of 100 to 200 miles depending upon the assumptions, as to the interior conditions, which were used in the computation.

Amplitude of the Vibrations. From the disastrous effects of earthquake waves it might be inferred that the vibrations have large amplitudes; but it is chiefly the suddenness of the vibrations that makes them effective. A sudden shock with an amplitude of $\frac{1}{5}$ inch is sufficient to shatter a chimney. It is the oscillation of the rock particles transmitting the vibrations that is here meant, not the movement of objects on the surface, which is often much greater. Just as a marble lying on a floor may be made to bound several inches by a slight tap of a hammer some distance away, so a sufficiently sudden rise of the earth-surface, for a fraction of an inch only, may project loose bodies many feet.

Destructive Effects. The disastrous effects of earthquake shocks result from (1) the suddenness and strength of rather minute vibrations of earth-matter, and from (2) the freedom of motion of the bodies affected. The deeper rocks, where sensibly continuous, transmit seismic vibrations without appreciable disruptive effect, so far as known, but bodies at the surface are fractured, overturned, and notably displaced. The tap of a hammer sends an almost imperceptible vibration along the floor, but this vibration throws well into the air the glass ball beneath which it runs. Similarly the minute seismic vibrations travel miles from their origin through continuous substance with little result, and may nevertheless then hurl a loose or unstable body to destruction. Rhythmical vibrations cause structures to sway more and more till they may topple over. Earthquake waves striking the sea-border may thrust the waters off-shore, and the return wave may overwhelm the coast (Fig. 293). The impact of the original shock results in great waves of translation, which are known to travel with the tremendous speed of 300 to 500 miles per hour; some having crossed the Pacific from Japan in about 12 hours. They are known by the Japanese name of *tsuna-*

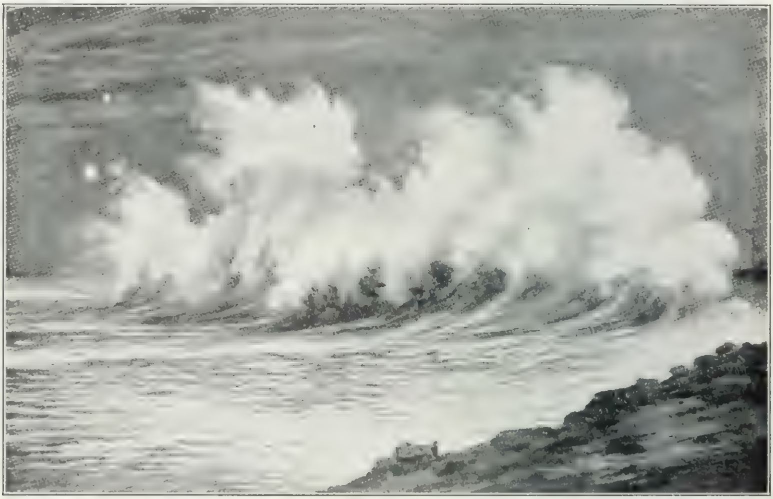
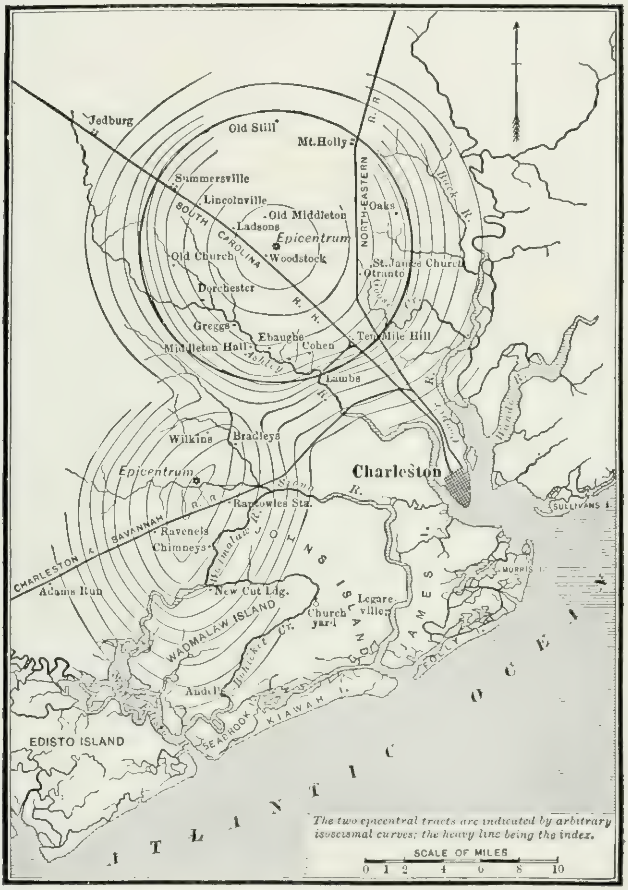


Fig. 293. — Great sea wave, tsunamis, on the coast of Ceylon. (Sieberg.)

mis. A large number, if not the majority of them, arise from submarine earthquakes.

Direction of Throw. At the *epicentrum*, which is the location on the surface directly above the subterranean place of origin, bodies are projected upwards. The destructiveness commonly increases for a certain distance from the epicentrum, and then diminishes. Lines drawn through points of equal effect (*isoseismals*) (Fig. 294) depart from circles or


Jedburg
Old Still
Mt. Holly
Summersville
Lincolnville
Old Middleton
Ladsons
Epicentrum
Woodstock
Old Church
SOUTH CAROLINA R. R.
NORTH-EASTERN R. R.
Oaks
Back R.
St. James Church
Otranto
Dorchester
Greggs
Ebaughs
Cohen
Ten Mile Hill
Middleton Hall
Ashley R.
Lambs
Cooper R.
Wando R.
Wilkins
Bradleys
Epicentrum
Charleston
Stono R.
Rantowles Sta.
SULLIVANS I.
CHARLESTON & SAVANNAH R. R.
Ravenels
Chimneys
Wadmalaw R.
JOHNS ISLAND
MORRIS I.
Adams Run
New Cut Ldg.
Legareville
Church yard
JAMES I.
WADMALAW ISLAND
Bohicket Cr.
FOLLY I.
Andells
SEABROOK I.
KIAWAH I.
EDISTO ISLAND
ATLANTIC OCEAN
SCALE OF MILES
0 1 2 4 6 8 10

The two epicentral tracts are indicated by arbitrary isoseismal curves; the heavy line being the index.

Fig. 294. — Epicentral tracts of the Charleston earthquake, with isoseismal lines. (Dutton, U. S. Geol. Surv.)

ellipses because of varying elasticity, density, etc., of the rocks through which the vibrations travel. As most earthquakes originate from slipping along fault planes, and the amount of slip varies from place to place, the original differences in the intensity of vibration at different points on the planes of origin, introduce inequalities in propagation and in surface effects.

Rate of Propagation. The rapidity of propagation of a seismic wave varies appreciably. The violent vibrations on the surface near the place of origin are the most irregular, but as the waves travel outward from the center of disturbance they become "sorted" out or separated into three more or less distinct sets of vibrations. This separation takes place because each set has a different speed and the faster ones run ahead of the slower ones. Each set also itself proceeds at different speeds depending upon the density and elasticity of the rock — greater speed being attained in less dense and more elastic rock. Two sets of waves spread through the body of the earth. The faster of these are the longitudinal waves, with speeds, in round numbers, of about 6 miles per second. They are followed by the transverse waves, with speeds of something like 4 miles per second. The third set of vibrations travels around the surface of the lithosphere with a speed of about 2 miles per second. The

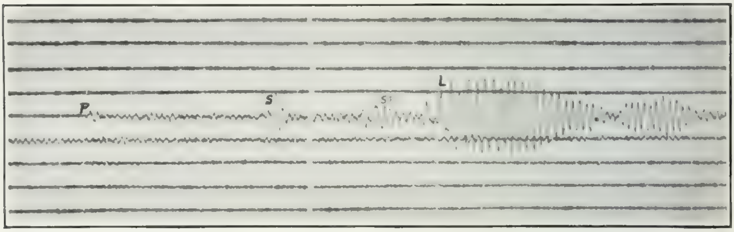


Fig. 295. — Seismogram of New Hebrides earthquake of March 22, 1925, recorded by Coast and Geodetic Survey instrument at Honolulu, Hawaii, approximately 3,970 miles away. P = arrival of fastest wave (longitudinal); S = arrival of second wave (transverse); L = arrival of long waves (having followed around the surface).

transverse waves which penetrate the earth are especially significant, for such waves *cannot pass through liquids* but are transmitted only by elastico-rigid material. They indicate that the bulk of the interior of the earth is solid and not liquid. While they have not yet been found to pass through a central core of the earth, having approximately one-eighth of the earth's volume, it remains for the seismologist to determine whether their apparent absence is due to loss by reflection and refraction, or to absorption by material of unknown character.

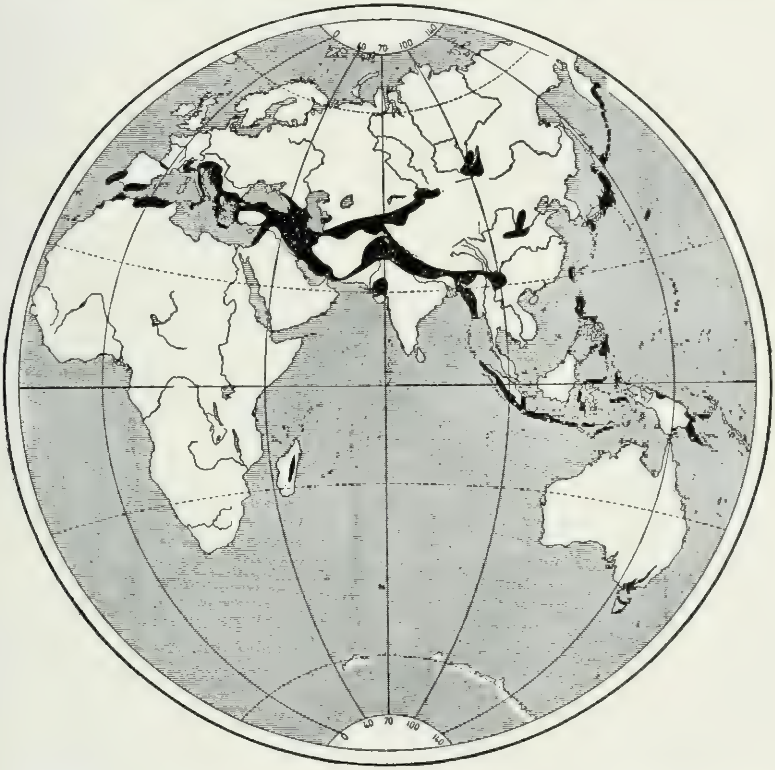


Fig. 296. — Map showing in black the principal earthquake regions of the Old World. (Montessus de Ballore.)

Distribution of Earthquakes. Over large portions of the globe, severe earthquakes are rare, but in certain regions they are, unfortunately, frequent. In general, earthquakes are likely to be rather frequent where geological changes are in rapid progress, as along belts of young mountains where the stresses are not yet adjusted and faulting is actively in progress, or on the great slopes, particularly submarine slopes, where adjustments to inequalities of stress are taking place, or at the mouths of great streams where deltas are accumulating, or about volcanoes where temperatures and strains are changing. In general the earthquake regions of the globe (Figs. 296 and 297) are also the regions of young

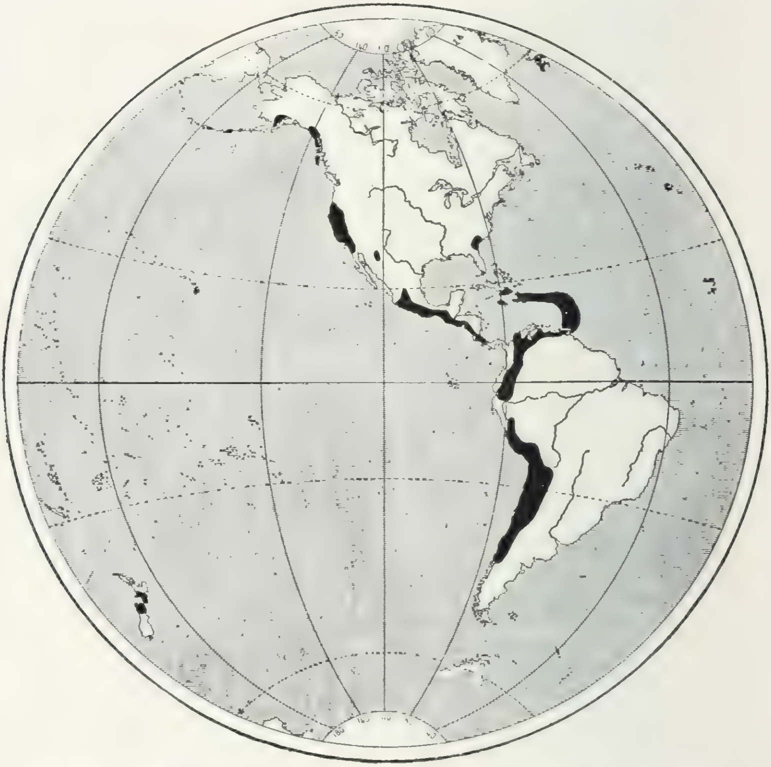


Fig. 297. — Map showing the principal earthquake regions of the New World. (Montessus de Ballore.)

mountains and volcanoes, which is natural, for these phenomena are genetically related.

The Geological Effects of Earthquakes

Geologically, earthquakes are of less importance than many gentler movements and activities. Disastrous as they sometimes are to human affairs, they leave few distinct marks which are more than temporary. Aside from the shaking down of loose material and local cracking in soft ground, they are of little or no geologic importance. Fracturing and minor faulting, which have sometimes been interpreted as the result of earthquake vibrations, are more often due to secondary adjustments arising from the major faulting. During the California earthquake of 1906, the ground was much broken along the line of the fault which caused the shock (Fig. 298).

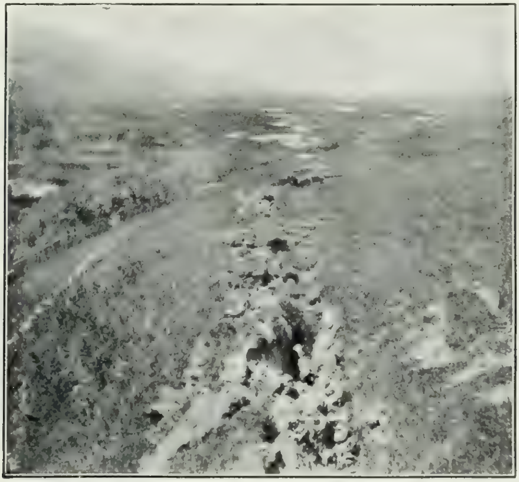


Fig. 298. — Characteristic surface appearance of the fault line, south end of Tomales Bay, Calif. (Newsom.)

Earthquakes sometimes dislodge masses of rock in unstable positions, as on cliffs and mountain sides, and occasion slumps and landslides. In 1920 in Kansu, China, a severe earthquake shock caused enormous masses of loess, made mobile by ground-water, to slide down the mountain sides, destroying several villages and damming small streams into lakes.

Effects on Drainage. The fracturing of the rock may interfere with the movement of ground-water. After new cracks are developed or old ones opened or closed, the movement of ground-water adapts itself to new conditions. It follows that springs sometimes cease to flow after an earthquake, while new ones break out. A strong flow of water issuing for a short time along a new avenue of escape, may wash up considerable quantities of sand if the surface material is incoherent. Craterlets and sand-cones are developed in this way (Fig. 299). Some excellent examples formed during the Charleston, S. C., earthquake in 1886.

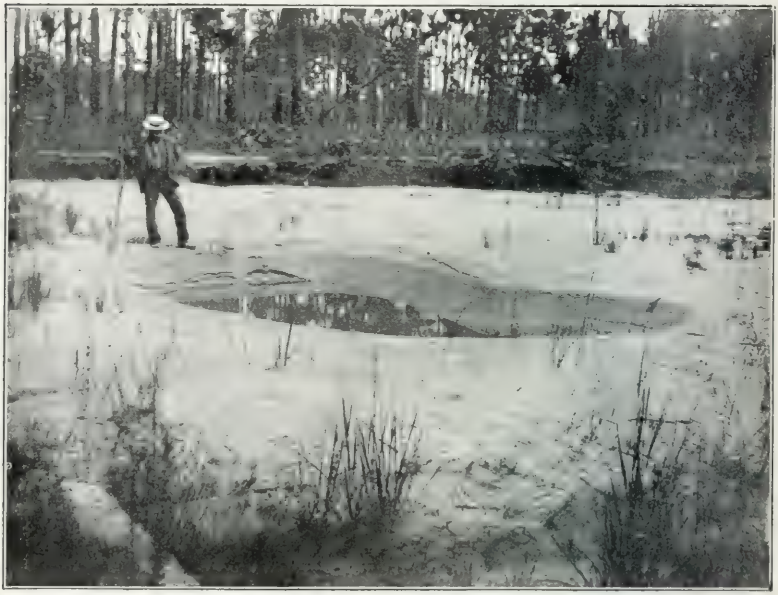


Fig. 299. — Craterlet formed during the Charleston earthquake. (U. S. Geol. Surv.)

In some parts of Yokohama, during the destructive earthquake of 1923, considerable areas of the city were flooded 2 and 3 feet deep by ground-water. The character of the water of springs is sometimes changed, presumably because it comes from different sources after the earthquake. Joints may be so widened as to intercept rivulets. Where faults accompany earthquakes they may occasion ponds or falls where they cross streams.

Effects on Standing Water. Some of the most destructive effects of earthquakes are felt along shores. The great tsunamis of the Lisbon earthquake (1755) and those accompanying the earthquake on the coast of Ecuador and Peru in 1868, were most destructive. Such waves, erroneously called tidal waves, have been known to advance on the land as walls of water 60 feet in height. They are most destructive on low coasts where the water may sweep over wide areas of land. The great loss of life during an earthquake is often caused by such waves. Lakes are also affected by earthquakes, but their waves are much feebler than those of the sea and are not often destructive.

Earthquake shocks are sometimes remarkably destructive to the life of lakes and seas. Thus during the Indian earthquake of 1897, "fishes were killed in myriads as by the explosion of a dynamite cartridge. . . and for days after the earthquake the river (Sumesari) was choked with thousands of dead fish . . and two floating carcasses of Gangetic dolphins were seen which had been killed by the shock."[1] This wholesale destruction of life is of interest since the surfaces of layers of rock, often of great age, are in places covered with fossils in such numbers as to indicate that the animals were killed suddenly and in great numbers, and their bodies quickly buried. It has been suggested that such rock surfaces may perhaps be memorials of ancient earthquake shocks.

Changes of Level. Permanent changes of level are sometimes accompanied by earthquakes. Thus a rise of 3 or 4 feet on the coast of Chili was observed at the time of the earthquake of 1822. Similar results have occurred on the same coast at other times, and on other coasts at various times. Depression of the surface is perhaps an even more common cause of earthquakes than elevation. On the coast of India all except the higher parts of an area 60 square miles in extent was sunk below the sea during an earthquake in 1762. Widespread depression in the vicinity of the Mississippi in Missouri, Arkansas, Kentucky, and Tennessee accompanied earthquakes of 1811 and 1812. Some of the depressed areas were converted into marshes, while others became the sites of permanent lakes. Reelfoot Lake mainly in Tennessee, is an example.

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[1] R. D. Oldham, *Report on the Indian Earthquake of June 12, 1897,* p. 80.

Changes of level are not confined to the land. Submarine earthquake faulting has broken telegraph cables. Considerable trouble has been experienced near Zante, off the coast of Greece. In 1878 after a snapping of the cable, the bottom was found to be so "irregular and uneven for a distance of about two miles that a detour was made and the cable lengthened by five or six miles." In still another case (1885) the repairing vessel found a "difference of 1500 feet between the bow and stern soundings."[1] These records point to sea-bottom faulting on a large scale, but it must not be supposed that the slip which broke the cable in 1885 amounted to 1500 feet. Most of the height of that fault-scarp was undoubtedly developed before the cable was laid down. A further displacement of a few feet was sufficient to snap the cable.

Bearing of Seismic Waves on Theories of the Interior of the Earth

Of the two kinds of waves that penetrate the earth, as we have seen, one set is condensational and the other distortional. Near the focus the two kinds are not distinguished from each other, but beyond about 700 miles, the speedier can be distinguished from the slower. Distortional waves which will not pass through a liquid are found to penetrate the outer part of the earth down to a depth of about half its radius, indicating a prevailing elastic solidity of this outer part. Down to the half-way point, an increasing velocity of these waves, in spite of increased rock density indicates increasing rigidity of the earth. Beyond that depth the records of the distortional waves are uncertain. It is premature to draw positive inferences respecting the constitution of the innermost part of the earth. However, the evidence of nutation of the earth, the body tides of the lithosphere, and seismic waves all point to solidity with no evidence of liquidity.

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[1] Forster, *Seismology*, 1877, summarized in *Amer. Geol.*, vol. 3 (1889), p. 182.

CHAPTER IX

VULCANISM

The term vulcanism is here used to include all movements of lava, as well as certain phenomena connected with these movements. The rise of lava assumes two general phases — movements by which lavas reach the surface, giving rise to eruptive or volcanic phenomena; and movements by which lavas intrude themselves into the outer formations of the earth and congeal underground. The first give rise to *volcanic* rocks, and the second to *plutonic* rocks. The first are *extrusive;* the second, *intrusive.* The fundamental nature of the two phases of vulcanism is the same.

I. EXTRUSIONS

When molten rock is forced to the surface it gives rise to the most intense and impressive of all geological phenomena. The energies acquired in the interior under great compression here find sudden relief. Enclosed gases often expand with extreme violence, hurling portions of lava to great heights and shattering them into fragments, special forms of which constitute *pyroclastic* (fire-broken) material. Much of the explosive violence of volcanoes has been attributed to the contact of the hot rising lava with the ground-water, but the function of ground-water in the explosions has probably been exaggerated, for water original to the lava is now thought to be much more important.

There are two phases of extrusion, and at their extremes they are strongly contrasted. One is explosive ejection, often of great violence; the other simply a quiet out-welling of the lava. More or less closely related to these two phases of extrusion are two classes of conduits, the one, *restricted* *openings*, as pipes, ducts, or limited fissures from which a relatively small amount of lava is extruded, which, congealing near the orifice, forms cones; the other, *great fissures* out of which the lava pours in large volume, and from which it spreads over great tracts, not uncommonly in broad, thin sheets. There is no fundamental difference, however, between great fissure eruptions and the eruptions of restricted vents, and the two types blend. The extent to which lava spreads into thin sheets depends rather upon the mass and the fluidity of the lava than upon the form of the outlet. The stupendous outflows of certain geologic periods appear to have issued mainly from extended fissures.

a. Fissure Eruptions

The chief known fissure eruptions of recent times are the vast flows of Iceland; but at certain times in the past there have been prodigious outpourings of lava, flow following flow until formations thousands of feet thick and covering thousands of square miles, were built up. Such outpourings occurred in Idaho, Oregon, and Washington (Fig. 300), where about 200,000 square miles were covered with sheets of lava aggregating in places some 2000 feet in thickness.

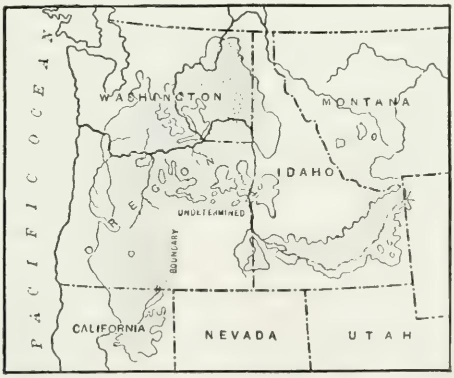

Fig. 300. — Lava flows of the northwestern part of the United States.

Still earlier, there were enormous flows on the Deccan in India covering a like area to the depth of 4000 to 6000 feet (Fig. 301). Still earlier, an even more remarkable succession of lava flows in the Lake Superior region developed a series of igneous rocks of almost incredible thickness. In these cases there is little evidence of explosive or other violent action.

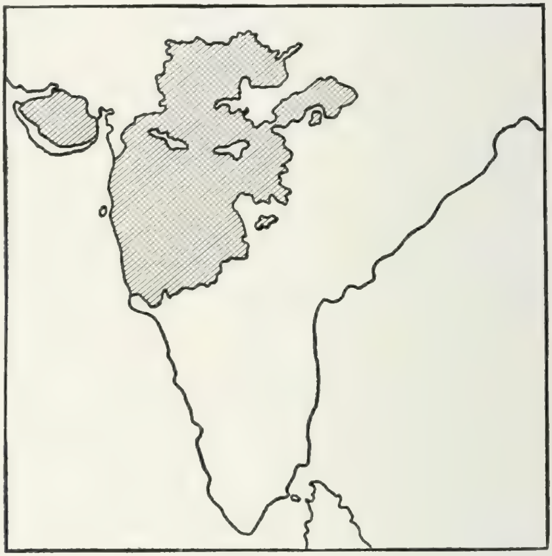

Fig. 301. — Flows of the Deccan. (After H. S. Washington.)

There are but few beds of dust, cinders, or other pyroclastic material. The inference is, therefore, that the lavas welled out quietly, and flowed over the surrounding country. For the most part these wide-spreading flows are composed of dark (basic) material, which is more easily liquefied and more fluent at a given temperature than the light (acidic) lavas. The latter are more disposed to form thick bodies near the point of extrusion.

It is notable that concomitant with the outpouring of great lava masses the region has a tendency to "founder," or sink, into the space previously occupied by the lava.

Diastrophic movements have, in many cases, subsequently raised the flows to plateaus.

Massive outflows were commonly not formed at one time; flow after flow occurred, gradually building the vast thicknesses. It is not uncommon to find sedimentary deposits, such as gravel, sand, mud, etc., between successive lava sheets, showing considerable lapses of time between flows.

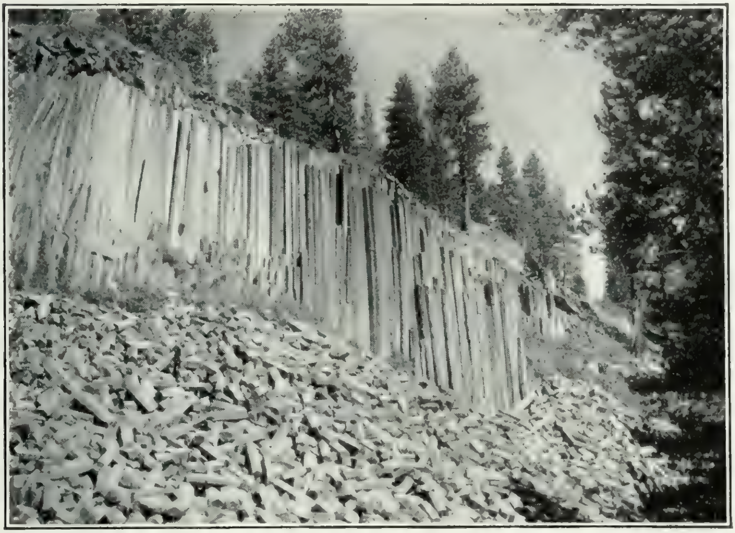

Fig. 302. — Columnar basalt. Devil's Post Pile. (Courtesy U. S. For. Serv.)

Successive flows are not usually co-extensive. If the later flows of the closing stages of a period of vulcanism fail to reach as far as the earlier ones did, a terraced or step-like aspect is given to the region, whence the name *trap-rock* (trappe, steps) is derived.

In many places, such lava sheets, especially if of basalt, assume, in cooling, a columnar structure (Fig. 302). The columns are rude six-sided prisms standing at right angles to the cooling surface. The explanation of columns is probably somewhat as follows: The surface of the lava contracts about equally in all directions on cooling. The contractile force may be thought of as centering about equidistant points. About a given point, the least number of cracks which will relieve the tension in all directions is three (*A*, Fig. 303). If these radiated symmetrically from a point, the angle between any two is 120°, the angle of the hexagonal prism. Similar radiating cracks from other centers complete the columns (*B*, Fig. 303). A five-sided column would arise from the failure of the cracks to develop about some one of the points (*C*, Fig. 303).

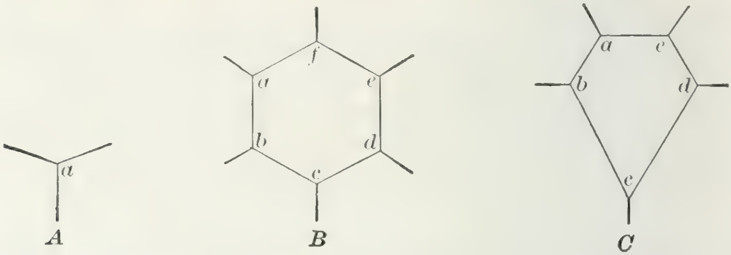

Fig. 303. — Diagrams to illustrate the formation of columns of basalt: A, the first stage in the development of a hexagonal column. B, the completion of a hexagonal column. C, a pentagonal column.

b. Volcanoes

A volcano is a roundish vent in the lithosphere, out of which issue gases and hot rock. The rock material is generally built up into mounds or cones (Figs. 304-306), some of which are large enough to make imposing mountain peaks. These cones are often called volcanoes, though in reality they are the result of volcanic activity. So long as the volcano is active there is likely to be a depression, or *crater* (Fig. 307), in or near the summit of its cone. Craters may be a mile or more across, but they are mostly smaller, some much smaller. While some cones and craters are beautifully symmetrical, others are irregular in shape, for eruptions from time to time break out in new places, destroying or modifying the old landmarks. Erosion of extinct volcanoes shows that the former passageways leading toward the sources of the lava vary much in size and shape, and usually have diameters smaller than those of the craters.

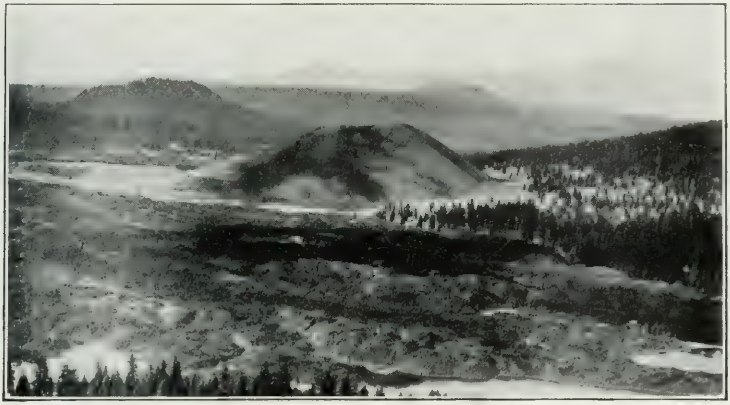

Fig. 304. — Cinder Cone. Lassen Peak, California, in background. (U. S. Army Air Serv.)

Number. The exact number of volcanoes now active cannot be stated with precision. In the first place, since most volcanoes are in action only periodically, it is impossible to

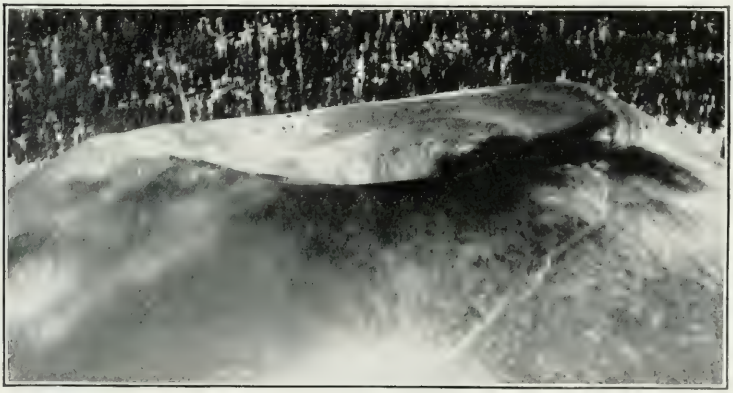

Fig. 305. — Crater of Cinder Cone, California. (U. S. Army Air Serv.)

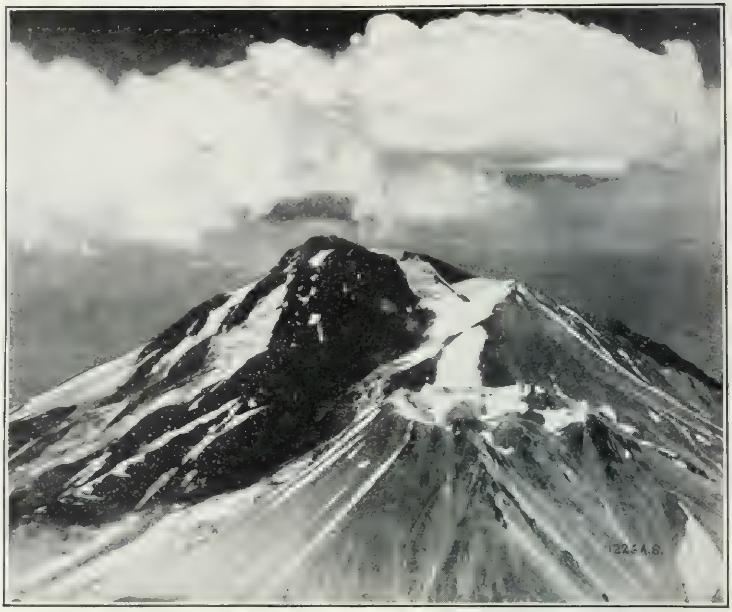

Fig. 306. — Mt. Shasta, with Shastina in foreground. (U. S. Army Air Serv.)

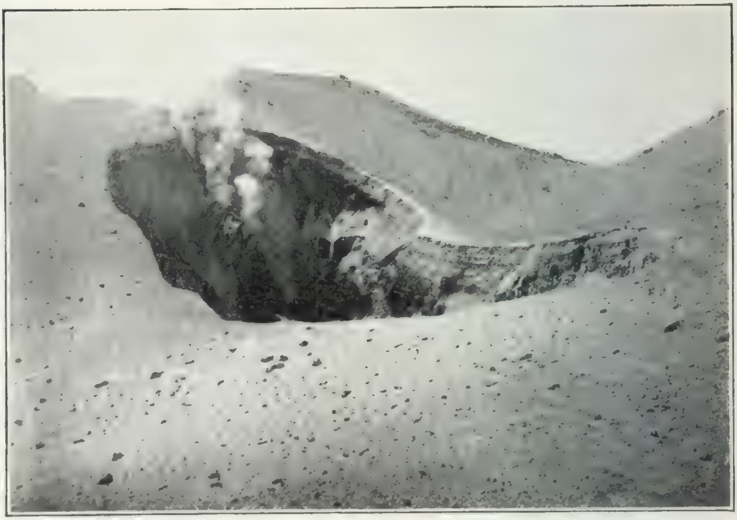

Fig. 307. — Crater of Lassen Peak, October 27, 1914. (P. J. Thompson.)

say whether a volcano that is now quiescent is extinct or only dormant; and in the second place, since volcanoes not uncommonly have subsidiary vents near them, it is often impossible to say whether two proximate vents should be considered one or two volcanoes. It is, however, quite safe to include at least 430 in the active list, and the number may be even more.[1] Of these 80 are submarine, having been reported by mariners. The number that cannot surely be called active, but which have been active so recently that their cones remain distinct, is several times as great.

Examples of Active Volcanoes

Many of the phenomena of volcanic activity may be understood from the description of a few typical vents.

Stromboli. Stromboli is a small island lying in the Mediterranean Sea forty miles north of Sicily. It consists of a cone rising over 3000 feet above the sea. From a vent about 1000 feet below the summit, steam is constantly issuing and, under proper atmospheric conditions, the steam is soon condensed into a cloud. Glow from red-hot lava so illuminates the under side of this cloud that it becomes visible at night from many miles distant and gives to this mountain the name "light-house of the Mediterranean."

During one visit the floor of the opening or crater was seen to be made of black rock. Traversing the floor were cracks through some of which steam issued with a puffing noise. In others lava was seen slowly bubbling. The bubbles rose, expanded to a foot or more in diameter, burst and subsided. In bursting many of the bubbles threw fragments of lava high into the air — even hundreds of feet. Many of these fragments fell back into the crater, others on the rim and slopes of the cone, thus building it up.

During another visit, several years later, five small active vents were observed. The most prominent and active one exploded at intervals with a sudden loud roar like the sound of a big cannon. Many red-hot, half-molten fragments were shot a thousand feet or more into the air, accompanied by a tall column of brownish fume. From a second vent, about thirty feet in diameter, came a shower of half-molten stones, puffs of white smoke, and a continuous roar. "The hole below was filled with thin, long, wavering, bright red flame-like tongues, which seemed to be spirits from the liquid lava not far below."[1] The other vents, though less active, showed the same phenomena.

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[1] Sapper, *Vulcanological Review*, III, p. 67, 1917.

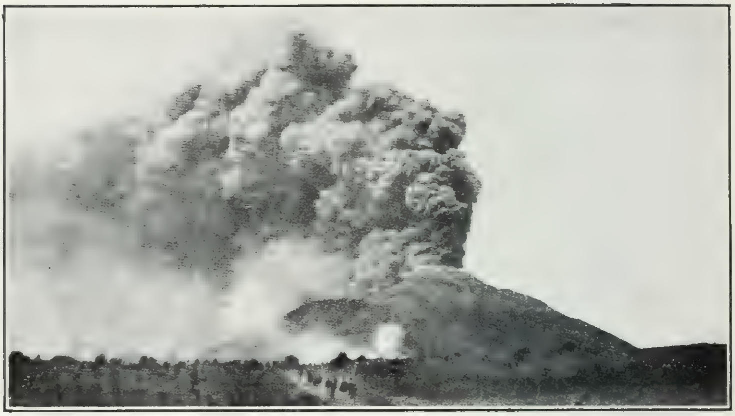

Fig. 308. — Vesuvius from the south, April 5, 1906. The white vapors are from lava that issued from near base of cone. (Perret, Carnegie Institution.)

At times of violent activity, when material is being ejected high and far, the noise of puffing, hissing, roaring, and exploding can be heard for miles. Stromboli is an example of the type of volcano that is almost constantly active, though it has brief periods of comparative repose.

Vesuvius. Vesuvius is a famous volcano whose cone rises about 4000 feet above the Bay of Naples and about ten miles southeast of the city of Naples. The present cone lies within the partly destroyed rim of a much larger crater. Records show that prior to 79 A.D. this large crater was wooded, even on its bottom, the volcano having been inactive for a long time. But in the year 79 A.D. a terrific explosion blew away a large part of the cone. The sky was darkened with this rock dust (popularly known as *volcanic ash*) which, as it fell on the surrounding country, buried and destroyed vegetation, houses, and even cities. Pompeii, a city of 20,000 inhabitants, was thus blanketed with falling dust, in places to depths of 25 to 30 feet. People and animals were killed, and buried where they lay. Apparently some lives were snuffed out instantaneously by hot, poisonous gases, for excavations have shown dinner guests still sitting at table under which a dog was lying curled as in sleep. On the other hand the upright position of a sentry still guarding the north gate of the city showed him to have remained courageously at his post and to have been buried as he stood. It is estimated that about 2000 people perished. Excavations have also revealed implements, utensils, works of art, and houses so well preserved that the culture of this Roman city is now well known.

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[1] H. S. Washington and A. L. Day, "Volcanoes of Southern Italy." *Bull Geol. Soc. Amer.*, Vol. 26 (1914), pp. 387-88.

Condensing steam gave rise to heavy rains during and after the eruption. The water falling on the dust produced streams of hot mud which flowed down the slopes of the mountain to engulf such vegetation and habitations as lay in their paths. The town of Herculaneum was overwhelmed and destroyed by mud flows probably over 50 feet deep at places.

Since then, Vesuvius has been periodically eruptive, with spells of quiescence between. In 1631, a specially violent eruption of steam and volcanic dust, followed by lava flows, killed some 18,000 people. Since 1700, there have been twelve important eruptive periods, during many of which there was great destruction of property and loss of life.

Krakatoa. The small island of Krakatoa, in the straits of Sunda, between Sumatra and Java, was known to be of volcanic origin, but for centuries there had been no noticeable activity. In the spring of 1883, symptoms of activity began to be noticed. There were earthquakes and deep rumblings, which gradually became stronger. A vent opened and dust was ejected into the air, making a pall of darkness for hundreds of miles around. Through the night of August 26th, the blackness was occasionally illuminated with flashes from the volcano, while at Batavia, one hundred miles distant, windows rattled and houses shook. At ten o'clock on the morning of August 27th three increasingly violent explosions culminated in so frightful a convulsion that the sound was heard 2200 miles distant in southern Australia and 3000 miles away in Rodriguez in the Indian Ocean. About two-thirds of the island was blown away, leaving the sea 1000 feet deep where this part of the island had stood. Part of the dust was projected to a height of 20 miles, where it was caught by the upper atmospheric circulation, and carried completely around the globe in about 15 days. Most of the broken rock, however, fell back into the adjoining sea, building the ocean floor up 10 to 12 feet for several miles around. The explosion produced air-waves which, according to barometric records, traveled three or more times around the globe. Great sea-waves were also generated, which swept across the intervening seas and rose 50 feet or more on the neighboring shores, destroying 295 villages and more than 36,000 people. These sea-waves traveled half-way around the earth. Like Stromboli and Vesuvius, the eruption was due to pent-up gases, chief of which was superheated steam.

Mont Pelée and La Soufrière. The volcano of Mont Pelée rises at the north end of Martinique, one of the volcanic islands of the West Indies (Fig. 309), as a cone, dropping steeply to the sea on all sides save the southwest. Here flat land along the coast was the site of the city of St. Pierre prior to the eruption of 1902. The crater was rimmed with wooded hills on all but the south side, where a notch appeared. For several months before the eruption, earthquakes had been noticed, dust had fallen, and by April 25th, poisonous sulphur gases had killed horses in the streets of the city. By May 6th Pelée was very active, belching out cinders and dust. All six of the telegraph cables from Martinique were broken on May 6th and 7th. Detonations like the fire of distant artillery were heard. On May 8th, the eruption reached its climax. Three terrific explosions shook the volcano. The second one, from a gash on the south flank of the crater, belched forth a tremendous cloud of hot gas and dust which swept down the side of the mountain with hurricane violence and two minutes later roared across St. Pierre. Houses were blown over, trees uprooted, statues torn from their pedestals and hurled many feet, and boats lying in the harbor were wrecked. The wreckage of the city

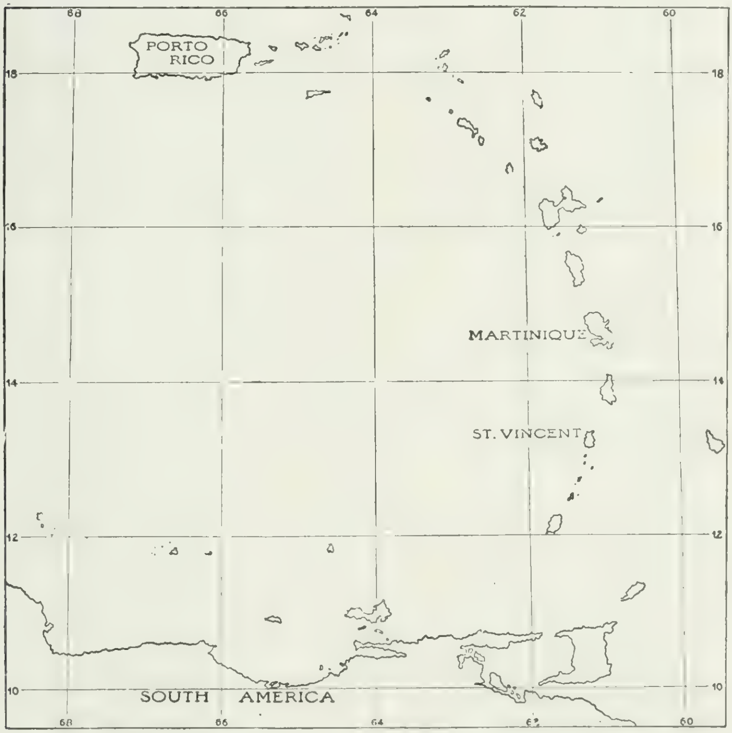

Fig. 309. — Map of the West Indies showing Martinique and St. Vincent.

burst into flame. Immediately following the destruction, a deluge of rain brought down the dust from the air in the form of mud, accompanied by a shower of pumice, plastering the

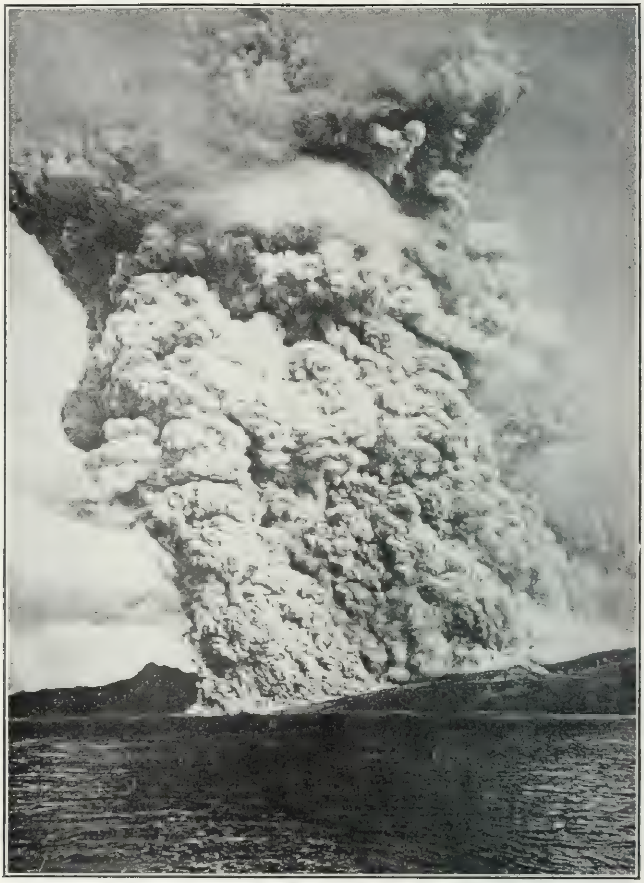

Fig. 310. — The eruption cloud of Pelée, December 16, 1902. (Lacroix.)

landscape, ruins, and victims with mud. With very few exceptions, the 30,000 people of the city were exterminated within a few minutes.

Several eruptions during the succeeding three months continued the devastation of the area with mud flows and dust falls. A cone of fragmental material was built within and overtopped the old crater rim. From within this cone a spire of solid rock, believed to be the plug of solidified lava, was pushed up to a height of 1210 feet above the crater rim. It soon crumbled away, however, and became inconspicuous.

La Soufrière, on the island of St. Vincent, 90 miles south of Martinique, also erupted on May 7th, 1902. For a year earthquakes and rumblings had gradually grown so that the people on the leeward side of the volcano had taken flight. On the windward side, however, the population remained, thinking that in case of an eruption they would be safe, and it was here that most of the 350 people who met death were killed by hot blasts, dust, and falling stones. The eruption continued for several days with abating violence. But on May 18th, another outbreak, of short duration, hurled stones so far that they fell in Kingston, 12 miles away.

Hawaiian Volcanoes. The island of Hawaii, so far as known, is built up practically entirely of volcanic material, from the bottom of the sea, 16,000 feet in depth, to the summits of Mauna Loa and Mauna Kea, each nearly 14,000 feet above sea-level, making these volcanoes perhaps the highest mountains of the earth. The crater of Mauna Loa, 3 miles long, 2 miles wide, and 1000 feet deep, is a very large one. When the volcano is quiet the flat floor of the crater is crusted over with solidified lava, but during an eruption it is raised, and parts of it become engulfed in lakes of molten lava, from the surface of which fountains may project liquid lava to heights of a hundred feet or more. Commonly the lava does not overflow the rim of the crater but issues through fissures in the side of the cone and flows sluggishly down the slope. Most of the flows solidify before going far, but some have reached the sea, 50 miles away. The lava assumes two distinct aspects: One is smooth and ropy (Fig. 311), containing little gas, and known as a *pahoehoe*, while the other is clinkery, fragmental, and highly gas-charged, known as *aa* (Fig. 312). As the lava flows out, the crater

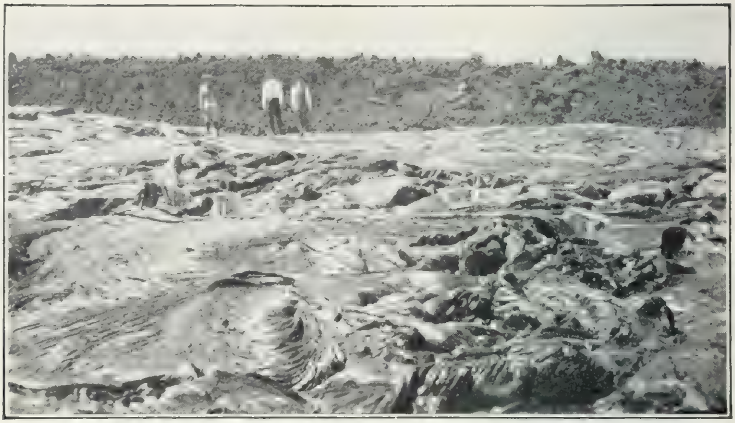

Fig. 311. — Ropy pahoehoe lava in foreground with clinkery aa lava in background. 1920 Kau Desert flow near Kilauea. (R. T. Chamberlin.)

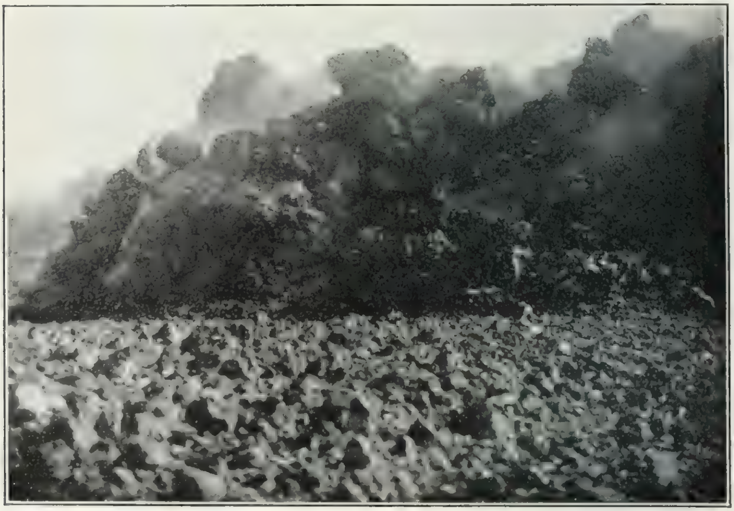

Fig. 312. — Front view of aa lava flow, Sakurajima volcano, Japan, January 18, 1914.

floor subsides, reaching its old position at the close of the eruption.

Kilauea. Situated twenty miles east-southeast of Mauna Loa, and 10,000 feet lower in altitude, is the active volcano of Kilauea. Its large crater, 9 miles in circumference, is, for the most part, crusted over with hardened lava, but in the southwestern part the pit of Halemaumau, more than 1000 feet across, usually contains molten lava (Fig. 313).

"The scene at Halemaumau presented to the visitor during the second week of November, 1920, as he looked down from the high rim, was a vast circular pit 1200 to 1400 feet in diameter, with nearly vertical walls, and at the bottom lay an inner landscape of smoking crags (Fig. 314), marginal floors reaching out to the walls, and pools of incandescent lava, large and small, the largest an octopus affair with its tentacles forming a fiery stream which swept in a torrent from one side of the bottom to the other. The stream ended in a cascade forking about a ledge and falling in brilliant rapids to a lower pool, where the glowing melt surged and swirled with scintillating golden patterns (Fig. 315). The pool where the stream rose at the southwest source-vent was at least 15 feet higher than the southeast sinkhole

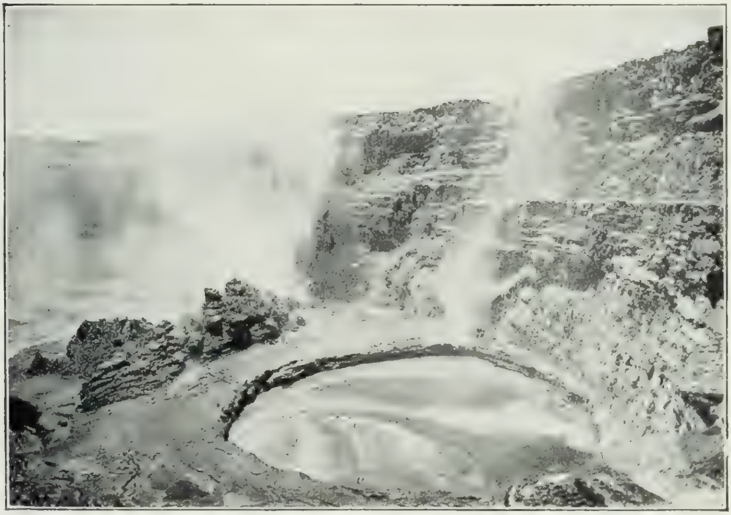

Fig. 313. — Halemaumau, September 7, 1920. (Jaggar.)

pool, where it ended in a fall. All of this flowing is a part of the convectional circulation whereby the hotter and more foamy lava rises at the west and the cooled and heavier substance sinks away at the east. The underground tubes

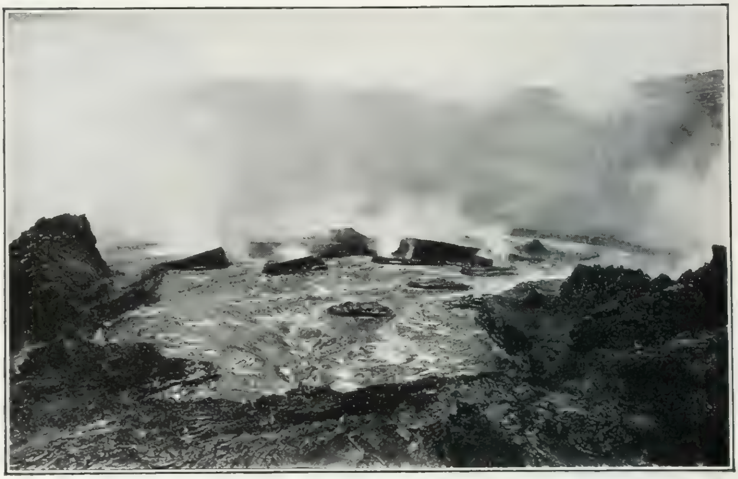

Fig. 314. — Halemaumau, September 7, 1920. (Jaggar.)

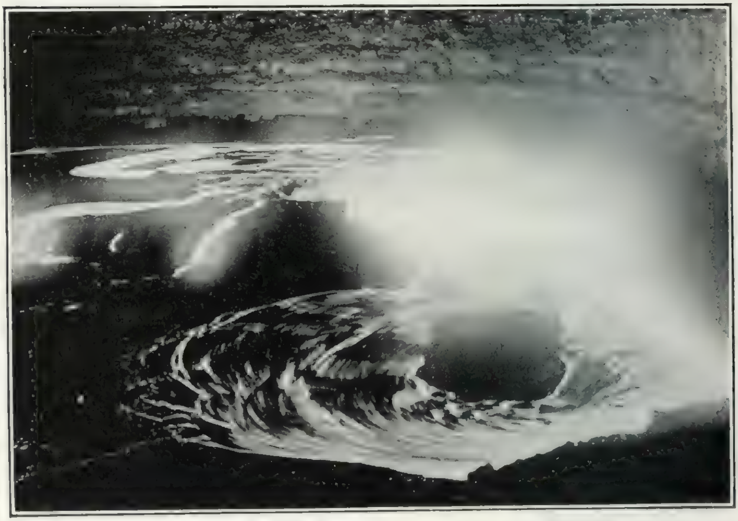

Fig. 315. — Halemaumau at night, May 12, 1923.

which convey the effervescing fluid ramify through the paste that forms the crags and floors, which paste extends indefinitely downward."[1] The whole floor of Halemaumau rises and sinks from time to time, the crag lava gaining on the lake lava at times, and anon the lake lava engulfing the crags. It is evident from long observation and from soundings that

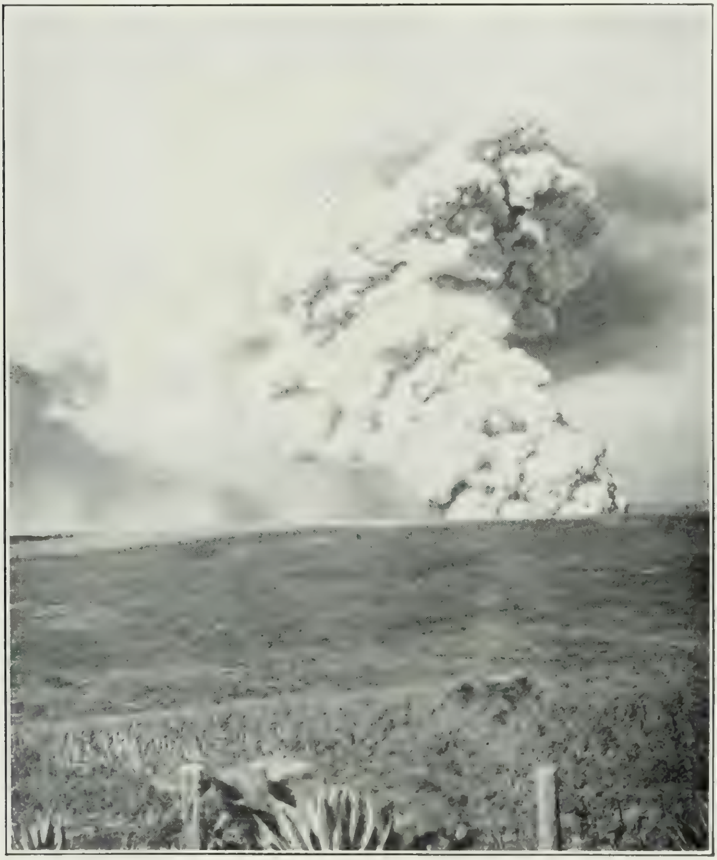

Fig. 316. — Cauliflower-like dust clouds from Halemaumau during the eruption of May 11, 1924. (Finch.)

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[1] T. A. Jagger, *Bull. Hawaiian Volcano Observatory*, Vol. VIII (1920), p. 129.

the lakes are shallow features existing for only a short time. "The whole process, with rise and fall of the complete lava column through a range of 600 feet, has been a marvelous demonstration of the *shallowness* of the *liquid* and the rigidity of the under substance on top of which the liquid is a foam, fed through the border cracks and through small wells generated along fissures."[1] Mauna Loa and Kilauea are largely independent in eruption, showing no hydrostatic equilibrium and hence no common reservoir. But in the type of sympathetic eruptions that is manifest between the two, they show the effects of a common cause of activity.

Phenomena Common to Volcanic Eruptions

Eruptions may differ in many ways but there are numerous features in common. Preliminary symptoms of activity are the rule. For days, months, and even years, rumblings and earthquakes are usually observed. The eruption itself consists of a pouring out of the upper part of the lava column or the blowing away of part of the confining cone, but is preëminently an emission of gas. Gas, therefore, may be considered the immediate heating and impelling agent in the eruption. If the gas escapes readily at the surface, a quiet type of eruption takes place, as in the Hawaiian volcanoes, where the lava boils, foams, wells up, and quietly flows from the crater. On the other hand, if the gas is confined, an explosion takes place, as at Krakatoa and Katmai. Commonly after a period of eruption, a period of quiescence ensues, during which more gas accumulates.

Distribution

1. In Time. In the earliest known ages (Archeozoic Era), igneous action appears to have been very widespread, though on account of the great metamorphic changes which these very old rocks have undergone, it is not always possible to determine their original nature. From the Paleozoic Era to the present the distribution of volcanic action over the surface seems to have been, in a general way, much what it is to-day; that is, certain areas were affected at times by volcanoes, while other and larger areas had few or none. However, in certain periods, there was widespread and energetic volcanic activity, while in other periods the same places had little or none. The known facts do not indicate a steady decline in volcanic activity, but rather a periodicity. One of the greatest of the volcanic periods falls within the Cenozoic Era, just preceding the present geologic period, and the volcanic activity of the present is perhaps but a declining phase of the activity of that time.

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[1] *Op. cit.*, p. 34.

2. Relative to Land and Sea. The active volcanoes of the present time are located chiefly along the borders of the continents, and within the great oceanic basins (Fig. 317). On this account, the sea-water has often been supposed to have some causal connection with volcanic action, and the presence of chlorine in the volcanic gases has been urged in support of this view. Volcanoes, however, are not distributed so equally and exclusively about the several oceans as to give this conclusion force, though basins, as basins, probably favor vulcanism. Volcanoes are numerous within and around the Pacific, the largest of the oceans; but they are not especially abundant in or about the Atlantic, while they are numerous in and around the Mediterranean, a much smaller body of water. On the other hand, there are existing or very recent volcanoes in the interior of Asia, Africa, and America far from the nearest ocean. Their distribution in the geologic past has been analogous. The abundance of volcanoes near the borders of the continents is clearly more the result of other conditions than of any connection between volcanoes and surface waters.

3. In Belts. In general the active volcanoes are arranged in belts (Fig. 317). The most notable belt well-nigh encircles the Pacific Ocean. Another lies in the Mediterranean region,

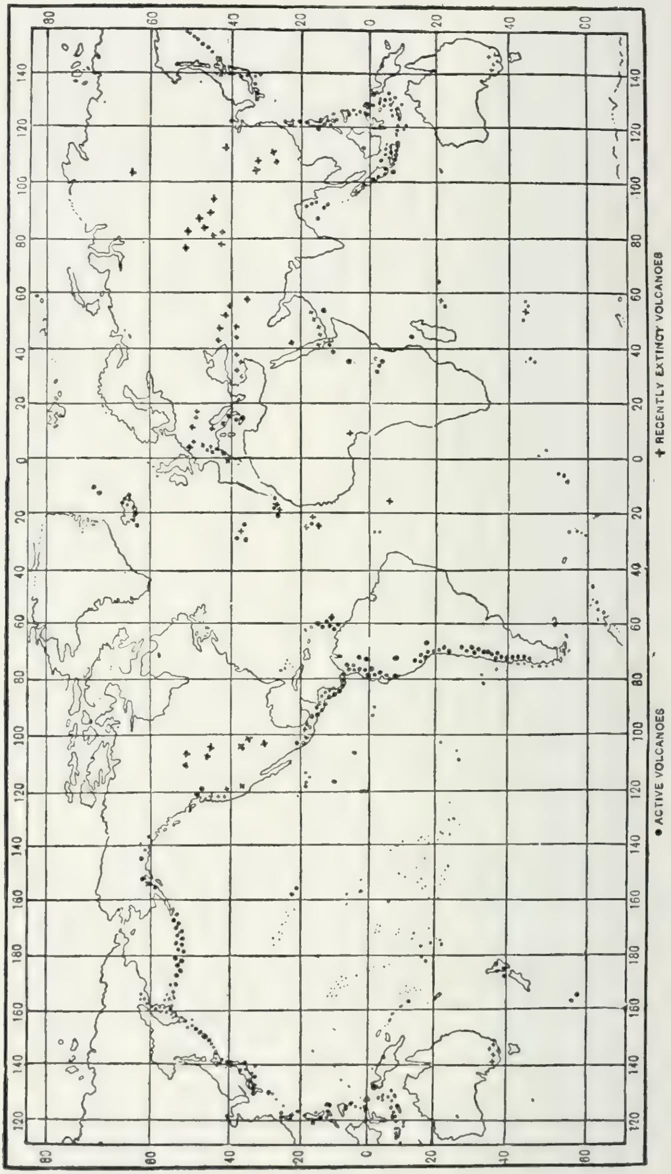
● ACTIVE VOLCANOES
✚ RECENTLY EXTINCT VOLCANOES
80
60
40
20
0
20
40
60
140
120
100
80
60
40
20
0
20
40
60
80
100
120
140
160
180
160
140
120

Fig. 317. — Map showing distribution of volcanoes. (After Russell.)

continuing far into Asia. In the Antilles, the Aleutian Islands, the Kurile Islands, and in some other tracts, there is a linear arrangement of volcanoes, with appreciable curvatures, the convexities of which are turned toward the adjacent ocean. In other cases there is a linear arrangement without appreciable curvature, as in the Hawaiian range and in western Italy from Bolsena to Vesuvius. There are many vents, however, which cannot be regarded as belonging to any belt.

4. Relative to Crustal Deformations. The distribution of present and recent volcanoes is much more suggestively associated with those portions of the crust *that have undergone diastrophic movement in comparatively recent times or are still moving.* The great mountain belt stretching from Cape Horn to Alaska and thence onward along the coast of Asia is dotted with active and recently extinct volcanoes. The tortuous zone of mountainous wrinkles that borders the Mediterranean and stretches thence eastward to the Polynesian Islands, is another notable volcanic tract. These two belts include the greater number of existing and recent volcanoes on the land. These are also the belts of diastrophic activity manifested in the distribution of earthquake foci (Figs. 296-297). All these are related phenomena, the crustal movements, whose manifestations are earthquakes and mountain chains, also facilitate the generation, migration and escape of lavas.

Products of Volcanoes

Pyroclastic (fire-broken) Material. The fragmental materials which are blown out of a volcano are, as a rule, portions of lava which solidified either before ejection, or during their flight through the air. Masses of rock tons in weight are sometimes thrown out, and from masses of such size the fragments grade down to minute particles of dust. The dust particles (often called ash), thrown high into the air in some cases and caught by the winds, are shifted incredible distances, as already noted. While, therefore, the fluid lava and the larger fragmental materials ejected from the volcano stay near the vent, the fine materials are scattered broadcast. Bits of lava may be ejected and solidify to pieces about the size of small gravel, known as *lapilli*. The coarser fragments are known as *cinders*, while those two, three or more inches in diameter are known as *bombs*.

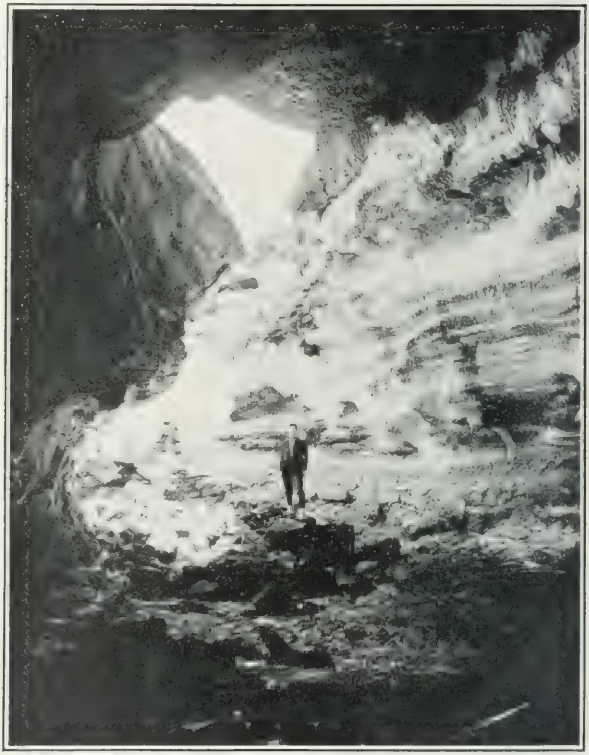

Fig. 318. — Postal Rift lava tube, September 7, 1920. (Jaggar.)

Liquid Rock, Lava. The term lava is applied to all kinds of liquid rock which issues from a volcano, and also to the solid rock formed when this congeals. Lava never flows so freely as water, and is sometimes very stiff or viscous. The distance to which it flows depends on its liquidity, its amount, and the slope of the surface on which it is poured. The more fluid the lava, the greater its amount, and the steeper the slope on which it flows, the farther it will move.

As lava flows, its upper surface may cool so much as to become hard while the interior is still fluid. The fluid part may then break out at the side or end of the hardened shell and flow away, leaving a hollow crust of solidified lava. Lava caves and tunnels are formed in this way. On further cooling the shell contracts and cracks, and often caves in, exposing the cavern or tunnel to view (Fig. 318). The hardened surface of a lava flow may be broken by the movement of the fluid lava below, and the solid fragments be displaced and upturned so as to give the surface a very jagged appearance. As it becomes solid, lava takes on various phases. Where it hardens quickly under little pressure, as at the surface, the gases and vapors which it contains may expand so that it is converted into a sort of solidified rock froth, called *scoria* (Fig. 319), or if the pores are very small, *pumice*. If certain kinds of lava solidify quickly, without becoming

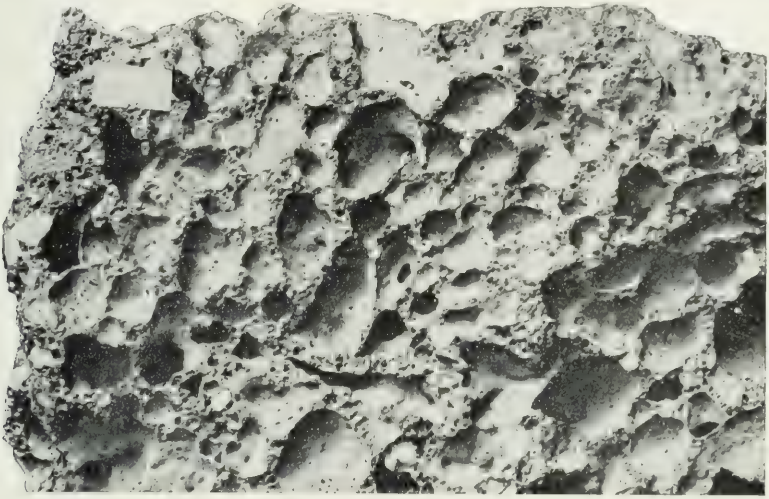

Fig. 319. — Scoriaceous texture. About $\frac{4}{5}$ natural size. (Church.)

frothy, they may make *volcanic glass*, or *obsidian*. If the lava cools slowly, the substances of which it is composed usually crystallize into various minerals. The kinds and proportions of the minerals depend chiefly upon the composition of the lava.

Gases. One of the most important features of volcanoes is the explosive action arising from the gases and vapors pent up in the lava. These occur partly in solution in the lava under high pressure, partly as compressed bubbles, while other portions are supplied afresh by chemical reactions as the conditions of temperature and pressure change. When crystallization of a lava sets in, a portion of the gas-content is held within the growing crystals, but another and more important portion is excluded from the crystals, and this excluded gas overcharges the remainder of the lava. If the gases are able to escape steadily from the liquid lava, and if not much crystallization takes place before the lava reaches the surface, the eruption is relatively quiet. If, however, the escape of the gases is retarded by the viscosity of the lava, and rapid crystallization of part of the lava supercharges the remainder of the liquid, gases gather in large vesicles or bubbles in the lava in the throat of the volcano, and on coming to the surface explode, hurling the enveloping lava upwards and outwards, often to great distances. In extreme cases, the pressure of the pent-up gases in the throat of a volcano may finally become so great as to blow away a large part of the crater walls or even the entire cone. Following this, as the upper portion of the lava column is removed, that below, relieved of much of the overlying pressure, explodes. This in turn relieves pressure from the lava next below, which also explodes, and so on down, until the supercharged part of the column has expended its force.

Steam is the chief volcanic gas. In most cases water vapor is exhaled from volcanoes in greater quantity than all other gases combined. From one of the minor vents of Mt. Etna, Fouqué estimated that 460,000,000 gallons of water in the form of steam were given off in 100 days. Among the other gases given off by volcanoes and lava flows are: carbon dioxide, carbon monoxide, sulphur dioxide, sulphur trioxide, hydrogen, nitrogen, hydrogen sulphide, methane, hydrochloric acid and hydrofluoric acid. Some of these are explosive at high temperatures when mixed with air in the proper proportions, and some are active poisons to plant and animal life. When igneous rocks are heated sufficiently in the laboratory, these same gases are given off in various proportions, indicating, among other things, that the solidified lavas have not lost all their gaseous constituents. The peculiar proportions of the rock-gases and volcanic gases seem to imply that they were not derived chiefly from surface waters or the atmosphere, but were original constituents of the rising lava. When given forth they appear to constitute real additions to the atmosphere. Given forth underground they play an important rôle in the process of contact metamorphism and in the genesis of ore deposits. The copper and zinc ore veins of Butte, Montana, and the gold-bearing veins of Tonopah, Nevada, are but two well-known illustrations of deposits arising from ascending *magmatic* waters.

Formation of Cones

Lava-cones. The lava usually flows away from the vent in streams, which solidify before running far. As the lava streams flow in different directions at different times, the total effect is a low cone formed of tongues of lava radiating from the point of exit. So far, therefore, as the volcanic cone is formed of lava, it has a radial structure, made up of a succession of congealed lava streams. In these cases the slopes are low, because the fluidity of the lava prevents the development of high gradients. It is, however, the exception, rather than the rule, that the cone is made up mainly of lava streams, though the great Hawaiian volcanoes are of this class. The form of the cone, when composed chiefly of lava, is also affected by the mass of the outflow and by its fluidity.

Other things being equal, the larger the outflow at a given time, the more widely it distributes itself and the flatter the cone.

Cinder-cones. The larger portion of the lava blown into the air by explosions falls back in the immediate vicinity of the vent and builds up a cinder-cone. From the nature of the case, this fragmental matter is often disposed symmetrically, making a cone with steep slopes (Fig. 304). Since the angle of repose for piled-up fragmental material is much higher than for flowing lavas, cinder-cones, on the average, are much steeper than lava-cones. Small or temporary vents formed as offshoots from the main vents often give rise to secondary or "*parasitic*" cones. On Etna, for example, these are numerous, and in certain other cases they are so important that the volcanic mountain is, in reality, a compound cone. A more subordinate variety consists of "*spatter-cones*" formed about small vents that eject little dabs of lava which form chimneys, cones, domes, etc. Spatter-cones (Fig. 320) often develop on the surface of the lava flows themselves.

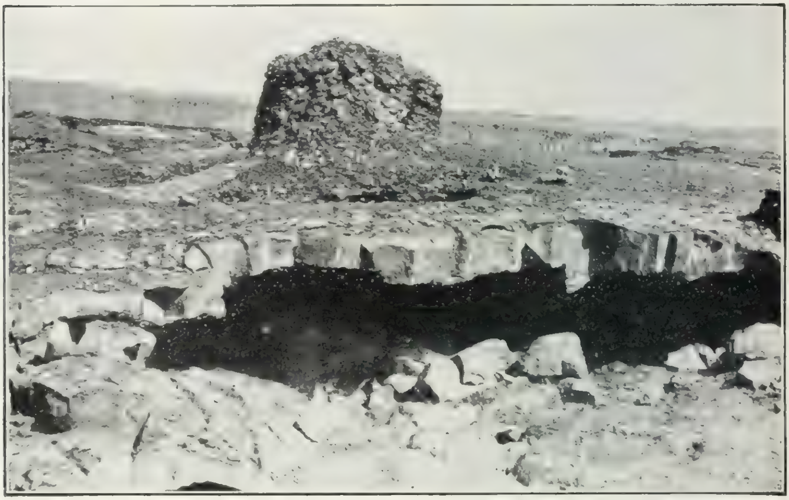

Fig. 320. — Spatter-cone and cavern. Kilauea, Hawaii. (Libbey.)

From most existing volcanoes both lava flows and fragmental ejecta are given forth, and the resulting cones are composite in material. The lava breaks through the side of the cone more frequently than it overflows the summit, and this produces irregularities of form and structure. Cones are also subject to partial destruction both by outbursts of lava and by explosions. As a result, many volcanic regions show old, partially destroyed craters, upon the ruins of which new and more perfect ones have been built.

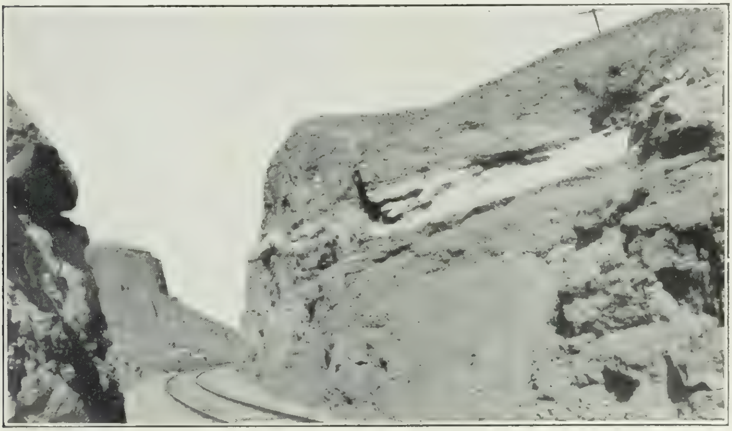

Fig. 321. — Beds of white volcanic dust and tuff interbedded with conglomerate. (Noble, U. S. Geol. Surv.)

In violent eruptions, the steam, accompanied with much dust, is shot up to great heights, often rolling outwards in cumulus or cauliflower-like forms (Fig. 317). The steam, by reason of its great expansion as it rises, and by its contact with the colder air, is condensed quickly, and prodigious floods of rain frequently accompany an eruption. This rain, carrying down a portion of the dust and gathering up much that has previously fallen, gives rise to *mud flows*, which in some cases constitute a large part of the final deposit. These mud flows lodge chiefly on the lower slopes of the cone or adjacent to its base. A portion of the finer exploded material floats away in the air to greater or less distances, and forms widespread deposits of volcanic dust, which, if later cemented into rock, constitute *tuff-deposits* (Fig. 321).

II. INTRUSIONS

Much of the lava which rises from the deep interior of the earth does not reach the surface, but solidifies below ground to form *intrusions*. The total amount which has risen toward, but not to, the surface, far exceeds all that which has flowed out at the surface. Intrusions are usually seen only after erosion has removed the rocks which overlay them.

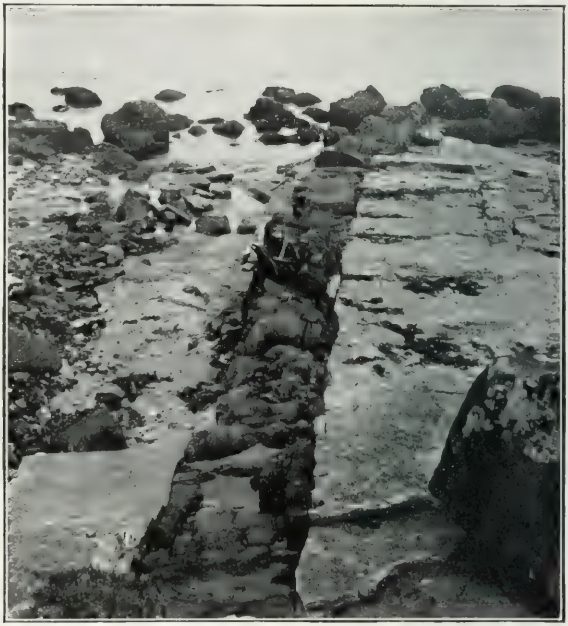

Fig. 322. — A dike two feet wide, cutting through sandstone. The exposed part of the dike is on land scarcely above the water. Arran, coast of Scotland. (H. M. Geol. Surv.)

If intruded lava fills a crack in the earth, it forms a *dike* (Fig. 322). Dikes may vary in width from a fraction of an inch to a hundred or more feet. If the lava is injected between layers of rock, it takes the form of *intruded sheets*, or *sills*. Sills likewise may vary in thickness from a fraction of an inch to hundreds of feet. If the lava accumulates below the surface in a large mass, arching up the overlying beds of rock, it forms a *laccolith* (lake of stone, Fig. 324). Laccoliths, varying in size from several hundred feet to several miles in diameter, are commonly circular or oval in ground-plan. If intruded lava solidifies in great masses, less regular in shape, and characteristically not arching the overlying rocks, *batholiths* are formed. The widely prevalent notion that batholiths have no "floor" beneath is a misconception. The majority of what have been called batholiths are probably flattish masses like enormous sills, or are tongue-shaped injected at a low angle, or roughly tack-shaped, in which the shaft of the tack is the solidified plug, and the head of the tack the spread-out portion near the surface. The spread-out portion may be many square miles in horizontal extent, though much more limited vertically. Laccoliths arch the overlying rocks by their own force of intrusion; batholiths are commonly secondary to mountain folding which prepares the way for the intrusion. Consequently the spread-out portions of typical batholiths tend to be longest in the same direction as the mountain folds or, if the mountains have been destroyed by erosion, par-

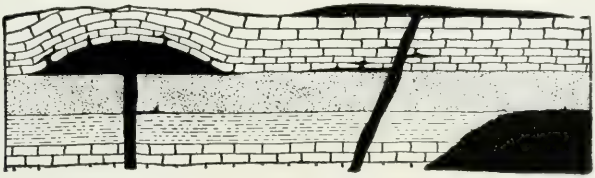

Fig. 323. — Forms taken by igneous rock.

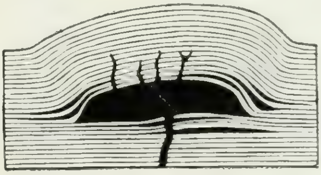

Fig. 324. — Ideal cross-section of a laccolith with accompanying sheet and dikes. (Gilbert, U. S. Geol. Surv.)

allel with the structural *grain* of the region. Batholiths may be large enough to form, after partial erosion, only single mountain peaks, or a chain of them may give rise to a whole mountain range. After the cone of an extinct volcano has been destroyed by erosion, the hardened lava which filled the old vent is known as a *plug*. Irregularly-shaped intrusive bodies intermediate between plugs and batholiths are called *stocks* or *bosses*.

Contact Phenomena. Where bodies of lava invade the country rock, the latter is commonly altered at and near the contact. It is, however, not always possible to distinguish between the effects of this *contact metamorphism* and that of dynamic or regional metamorphism effected by great earth movements. But in general there is (1) baking of the sediments, with the loss of water and carbon dioxide; (2) recrystallization into larger grains and coarser texture, and particularly the crystallization of such impurities as are present into minerals such as garnet, hornblende, tourmaline, etc., and (3) the addition of substances from the lava to form new minerals. This last process, known as *mineralization*, has given rise to some very important ore deposits. In the vicinity of large intrusive masses, the metamorphic effects may be traced for considerable distances—even miles—from the contact. Near small masses the zone of metamorphism commonly measures only a few inches or a few feet.

Products of Intrusion

When intrusions solidify, igneous rock is formed. There are many different kinds of igneous rock, the differences depending partly on the chemical composition of the lava, and partly upon the conditions of solidification. In order to understand igneous rocks it is necessary to know certain facts concerning the nature of lava.

Nature of Lava. We commonly think of lava as molten rock, but this view is not strictly correct. Lava is rather a solution of mineral matter in mineral matter. A simple analogy may illustrate what is meant. If pounded ice and salt are mixed together at a temperature of 30° F., the two form a liquid, though the temperature is too low to melt either one by itself. We commonly say the salt is dissolved, but it would be quite as correct to say that the ice is dissolved in the salt. The ice did not melt, because the temperature was too low to melt it. The two minerals, ice and salt, are dissolved in each other, and the solution takes place at a temperature below the melting point of either. Something of the same sort appears to take place when rock becomes liquid. The liquid form is assumed at a temperature below that which would be necessary to melt the minerals which make up the lava.

The propriety of regarding lava as a solution of minerals in minerals is shown in another way. Granite, an igneous rock, is made up chiefly of three minerals, quartz, feldspar, and mica. These minerals melt at different temperatures. If lava were nothing more than mineral matter melted, the several minerals should solidify from the liquid as it cools, *in the reverse order of their fusibility.* Of the above-mentioned minerals quartz, which has the highest melting temperature, should be the last to melt when heated; and if lava were merely melted rock, quartz should be the first to take solid form as the lava cools, and it should become solid when the lava reaches the freezing temperature of this mineral. But this is not what happens. Quartz, the least fusible mineral in the case of granite, is usually the last to take solid form. In other words, *the order in which the various minerals solidify from the liquid lava is independent of their fusibility.* This would not be the case if they were simply melted.

The liquid lava is essentially fluid glass. It is analogous even to common glass, except for the fact that manufactured glass is relatively free from iron and other coloring substances which abound in lavas, rendering them dark or more or less opaque. Lavas, too, are usually mixtures of several silicates, while manufactured glasses consist of only one, or at most only a few. Furnace slag is essentially an artificial lava.

Temperature of Lava. Accurate determinations of the temperatures in the middle of lava-columns, where they have been least reduced by contact with the rock walls, have not been made; but it is clear from the yellow heat of some lavas that their temperatures are often appreciably above their liquefaction-point. This is also a necessary inference from the length of time lavas remain fluid, notwithstanding their ascent through cooler rocks, the conversion into steam of the water in the rock through which lavas pass, and the expansion and escape of the gases. In some cases where determination has been practicable, it has been found that the melting-points of silver (about 960° C.) and copper (about 1060° C.) are reached. Careful measurements on Hawaiian lava give 750° – 850° C. and also bring out the striking fact that temperatures of the surface portion of the lava lakes are appreciably higher than those a score of feet below the surface. This is evidently because some heat is produced by chemical reactions among the gases and between the gases and certain constituents of the liquid as the lava approaches the surface and pressure is reduced. For example, the union of three gram-molecules of ferrous oxide and one of water gives one part of magnetite, two parts of hydrogen, and liberates 15,400 calories of heat.

Solidification and Crystallization. When a lava is cooled quickly, the dissolved silicates solidify in their diffused condition essentially as they were in the liquid, for there is not sufficient time for the molecules of one kind to come together in the systematic order necessary to form crystals. The essential feature of crystallization is this arrangement of the molecules of a given kind, according to a definite plan, giving a specific crystal form.

In a thick, viscous liquid, this systematic arrangement of molecules into definite crystal forms takes places slowly, for the crystallization force in silicates is far less energetic than in water, which crystallizes into ice rapidly and with great force. Because of the slowness of crystallization, the solidification of lava may catch the process of crystallization at any stage. If the lava is cooled quickly, the result is a glass; if less quickly, part glass and part crystals; if slowly enough, all becomes crystalline. While some other factors influence the final result, in general, *the slower the cooling, the larger the crystals.*

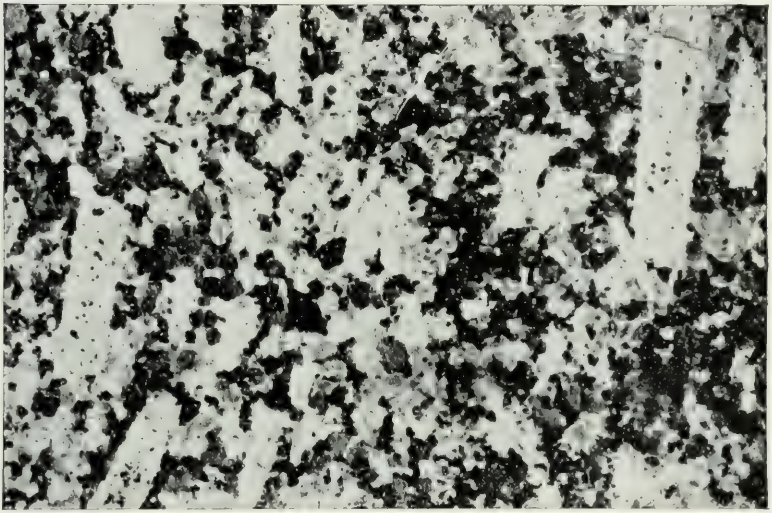

Fig. 325. — Coarse-grained rock (granite) about half natural size. The white patches represent crystals of one or two kinds of mineral, and the dark parts represent crystals of others.

The crystal forms of the different minerals are unlike. For example, a crystal of quartz, if nothing interferes with its growth, always takes the form of a six-sided prism terminated by pyramids (Fig. 331). The crystals of feldspar, mica, etc., take other forms (Fig. 329). All crystal forms are grouped into six (sometimes made seven) fundamental systems of crystallization, and there is a multitude of variations of special form in each. The treatment of these forms belongs to the special subject of crystallography. The first crystals to form in a liquid lava tend to assume perfect shapes, but when many crystals are forming at the same time, as is generally the case when lava solidifies, the crystals interfere with one another's growth. The result is that they interlock in all sorts of ways, and the forms of most of the crystals are very imperfect.

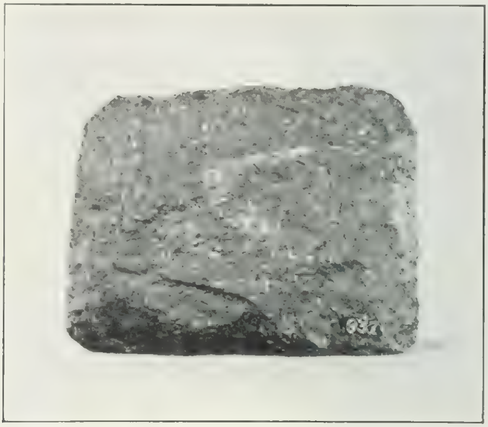

Fig. 326. — Fine-grained rock (basalt).

If the crystals are large enough to be distinct to the eye, even without close scrutiny, the rock is *coarse-grained* (Fig. 325). If the crystals are so small that they cannot be seen readily, the rock is *fine-grained* (Fig. 326).

Successive Stages of Crystallization. Since eruptions may take place intermittently, it is obvious that cooling of the lava may be in progress beneath the surface during the intervals between eruptions. After a certain stage of partial crystallization has been reached during such time of quiet, a renewal of eruption may take place and the whole mass of lava be shifted into new surroundings, with the result that a second phase of solidification may be superposed on the one already started. The rock may then show two phases of crystallization: (1) large crystals of the kind or kinds which developed in the lava during the first stage of slow subterranean cooling; and (2) small crystals or glass developed during the more rapid cooling under the new conditions. The result is large crystals (*phenocrysts*) set in a matrix of small crystals or of glass. Rock of this physical composition is *porphyry* (Fig. 327), a term which has a textural but not a mineralogical significance.

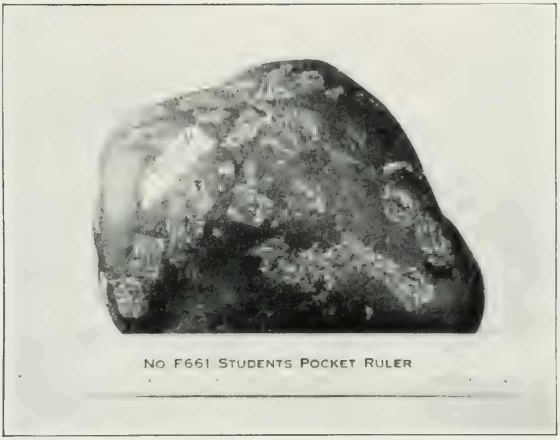

Fig. 327. — Porphyritic rock.

Composition of Igneous Rocks

Nearly all the chemical elements known on the earth are found in igneous rocks, though but few of them are abundant. These few are regarded as the normal or essential constituents, while the rarer substances are regarded as incidental. The relative amounts of the more abundant elements in the crust of the earth, as nearly as now known, are shown in the table on page 358:[1]

This table shows that only eight of the elements exceed one per cent, and only one other reaches one-half of one per cent. Many of the elements that are of the utmost importance in the affairs of men occur in quantities too small to

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[1] F. W. Clarke, *U. S. Geol. Survey, Bull. 770.* (1924).

Element	*Symbol*	*Percentage in the solid crust*
Oxygen	O	46.46
Silicon	Si	27.61
Aluminum	Al	8.07
Iron	Fe	5.06
Calcium	Ca	3.64
Sodium	Na	2.75
Potassium	K	2.58
Magnesium	Mg	2.07
Titanium	Ti	.62
Hydrogen	H	.14
Phosphorus	P	.12
Carbon	C	.09
Manganese	Mn	.09
Sulphur	S	.06
Chlorine	Cl	.05

be estimated in percentages. The precious metals, and even some of the more common ones, as lead, zinc, and copper, are of little importance quantitatively. We do not know just how the various elements of igneous rocks are united in the liquid lava, but their combinations after the lava has become solid may be determined.

In a general study of igneous rocks we may for the present neglect all but the first eight of the elements. In the crystallization of lava to form igneous rocks, three steps may be considered: (1) the union of elements to make compounds; (2) the union of compounds to make minerals, and (3) the aggregation of minerals to form various kinds of rocks.

1. Combination of Elements. The union of oxygen with the other seven elements may be taken as a fundamental step in the series of combinations. The result is the following *oxides:* [1]

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[1] F. W. Clarke, and H. S. Washington, *Prof. Paper 127, U. S. Geol. Survey* (1924).

Oxide	*Symbol*	*Average Percentage in Igneous Rocks*
Silica	SiO_2	59.14
Alumina	Al_2O_3	15.34
Ferric Oxide	Fe_2O_3	3.08
Ferrous Oxide	FeO	3.80
Lime	CaO	5.08
Magnesia	MgO	3.49
Soda	Na_2O	3.84
Potash	K_2O	3.13

Oxygen sometimes unites in proportions different from those given here, (e.g., combination of ferrous and ferric oxides gives the magnetic oxide of iron, Fe_3O_4) but for the present discussion other exceptions may be neglected.

2. Formation of Minerals. One of these eight oxides, silica, acts as an acid, while the others act as basic oxides. This fact is so significant that rocks containing more than 65% silica are called *acid rocks,* those with 55% to 65%, *intermediate rocks,* and those with less than 55%, *basic rocks.* The significance of this classification will appear shortly. Combination of the acid oxide (SiO_2) with the basic ones gives rise to silicates, which are the most abundant minerals of igneous rocks. The union of silica (SiO_2) and lime (CaO) gives calcium silicate (CaO,SiO_2) or $CaSiO_3$. The union of silica with magnesia gives magnesium silicate (MgO,SiO_2) or $MgSiO_3$. Corresponding combinations of silica with the other oxides named give rise to other silicates. These silicates occur in nature as minerals.

However, in the formation of silicate minerals, two additional principles are to be noted: (*a*) In the first place, silica unites in different ratios with the bases, thus giving rise to *sub-silicates, ortho-silicates, bi-silicates, tri-silicates,* or *poly-silicates.* While all the bases are not known to combine in all these different ways, many do so in more than one way. (*b*) In the second place, silica unites with two or more of the bases at a time, so that we may have a calcium-aluminum silicate, or calcium-magnesium-aluminum silicate. Not all of the possible combinations, however, actually exist as minerals.

Minerals other than silicates also occur as minor constituents of igneous rocks. Elements such as sulphur, phosphorus, chlorine, fluorine, etc., or their derived acids, unit with bases to form sulphides, sulphates, phosphates, chlorides, fluorides, etc. In addition, there are native metals in some igneous rocks. But all together these minor compounds hardly reach more than one or two per cent of the whole.

As a result of all this variety of combination, the total number of minerals in igneous rocks is large. Geology deals with these minerals as constituents of the earth, but those that are really essential to form igneous rocks are few. These few rock-forming minerals fall naturally into four groups.

3. Combination of Minerals into Igneous Rocks. Fortunately for the simplicity of geological study, a few minerals make up the great mass of igneous rocks. These few are *quartz*, the *feldspathic minerals*, the *ferromagnesian minerals*, and the *iron oxides*.

Quartz is the free acid already mentioned. If a magma contains more silica than is required to combine with the bases, the excess crystallizes as quartz. Crystals of quartz are typically six-sided prisms, capped by six-sided pyramids (Fig. 331); but in igneous rocks their forms are usually very imperfect. Some of the leading characteristics of quartz are given on page 367.

The *feldspathic group* of minerals embraces those formed by silica in union with alumina, together with either potash, soda, or lime, or two or more of these together. The feldspathic minerals are normally light in color, ranging from white to red, gray or greenish. The feldspars are minerals of the first importance in igneous rocks. The varieties of feldspar are given on page 364.

The *ferromagnesian* minerals are those formed by the union of silica with iron, magnesia, and lime, together with more or less of the other basic oxides. The ferromagnesian minerals are normally dark (commonly blackish and greenish) from the presence of iron, the great coloring element of rocks, but the color distinctions do not hold in detail, and cannot be safely trusted as a means of identification. Among the ferromagnesian minerals the most important are the *pyroxenes*, the *amphiboles*, and the *micas*. The pyroxenes and amphiboles have nearly the same chemical composition, but differ in crystal form and physical properties. They are among the most abundant of the dark minerals in crystalline rocks.

The two leading *micas* are the iron-magnesia mica, *biotite*, or black mica, and the potash mica, *muscovite*, or white mica, the familiar "isinglass" of the stove-door. Chemically, muscovite should go with the potash feldspars, but it is distinguished from them by its crystalline habit and physical properties. Chemically, biotite should go with the pyroxenes and amphiboles, which it closely resembles except in the form and properties of its crystals.

Two *iron oxides*, magnetite (Fe_3O_4) and hematite (Fe_2O_3), are widely disseminated in igneous rocks, but they are far less abundant than the silicates. They constitute the free bases already mentioned.

Summary of Salient Facts

The salient facts are, therefore, (1) that out of the known ninety chemical elements of the earth, eight form the chief part of its exterior; (2) that one of these elements uniting with the rest forms eight leading oxides; (3) that one of these oxides acts as an acid and the rest as bases; (4) that by their combination they form a series of silicates of which a few are easily chief; (5) that these silicates crystallize into a multitude of minerals of which again a few are chief; and (6) that these minerals are aggregated in various ways to form igneous rocks. Possessed of these leading ideas, we are prepared to consider the different kinds of igneous rocks.

REFERENCE LIST OF THE MORE COMMON ROCK-FORMING MINERALS, AND A FEW OF ECONOMIC IMPORTANCE

Actinolite. A magnesium-calcium-iron amphibole; commonly bright green to grayish green; crystals usually slender or fibrous.

Agate. A banded or variegated chalcedonic quartz.

Alabaster. A fine-grained variety of gypsum, either white or delicately colored.

Albite. A soda feldspar, an aluminum-sodium silicate; H.[1] 5–6; cleavage perfect in two planes; luster vitreous or pearly white; occasionally bluish gray, reddish, greenish; sometimes opalescent.

Amethyst. A variety of quartz or corundum of purple or bluish violet color, due probably to manganese.

Amphibole. The type of an important group of rock-forming minerals known as the amphibole or hornblende group; a ferromagnesian silicate; monoclinic; H. 5–6; luster vitreous to pearly; fibrous varieties often silky; black, ranging through various shades of green to light colors; embraces the magnesium-calcium varieties, tremolite and nephrite; the magnesium-calcium-iron variety actinolite; the aluminous-magnesium-iron-calcium variety *hornblende*, the most important member of the group, and others.

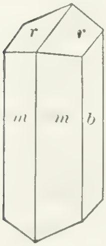

Fig. 328.—Amphibole crystal.

Andesine. A plagioclase feldspar; a sodium-calcium-aluminum silicate, intermediate in composition between albite and anorthite; H. 5–6; white, gray, grayish, yellowish, flesh-red; luster subvitreous, inclining to pearly.

Andalusite. An aluminum silicate; luster vitreous; whitish, rose-red, flesh-red, variety pearly gray, reddish brown, olive-green; H. 7.5, infusible; impurities sometimes so arranged in the interior as to exhibit a colored, crossed, or tesselated appearance in cross-section (chiastolite.)

Anhydrite. Calcium sulphate; H. 3–3.5; luster pearly to vitreous; white, sometimes bluish or reddish; differs from gypsum in absence of water and in its greater hardness.

Anorthite. A plagioclase feldspar; a calcium-aluminum silicate; varies much by impurities and admixtures; H. 6–6.5; pearly or vitreous luster; white, grayish, reddish.

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[1] H. = Hardness.

Anthracite. Hard coal; supposed to be derived from bituminous coal by metamorphism.

Apatite. Essentially calcium phosphate with chlorine or fluorine; hexagonal; H. 5; luster vitreous or sub-resinous colors, usually greenish to bluish, characterized by a hexagonal form.

Aragonite. Calcium carbonate; differs from calcite in *cleavage*, and in being orthorhombic; H. 3.5–4; luster vitreous or resinous; white, also gray, yellow, green, and violet.

Asphaltum. Asphalt; mineral pitch; a natural solution of different hydrocarbons; odor bituminous; melts at 90° to 100° C.; burns with a bright flame; graduates into mineral tars and through these into petroleum; probably the residue of the latter.

Augite. One of the pyroxenes; an aluminum-calcium-magnesium, iron silicate; H. 5–6; monoclinic, crystals usually thick and stout; sometimes lamellar; also granular; black, greenish black, deep green; an important rock-forming mineral.

Bauxite. Essentially hydrated alumina; occurs in concretionary grains of clay-like material, whitish to brown; valuable as a source of aluminum.

Barite. Barites, heavy spar, barium sulphate; orthorhombic, H. = 3–3.5; luster vitreous to resinous, sometimes pearly; white, inclining to yellow, gray, blue, red, or brown; very heavy, sp. gr. 4.3–4.7.

Biotite. Black mica, a potash-aluminum-magnesium-iron silicate; monoclinic; perfect basal cleavage into thin laminae; sometimes occurs as a massive aggregation of cleavable scales; H. 2.5–3; luster splendent on cleavage surface; black to dark green; cleavage surfaces smooth and shining; a very common constituent of crystalline rocks.

Bitumen. Includes asphaltum and also some more liquid compounds of the same general sort.

Calcite. Calcspar; calcium carbonate; rhombohedral, perfect rhombohedral cleavage; often taking the forms known as dogtooth spar, nail-head spar; frequently stalactitic and stalagmitic; H. 2.5–3.5; luster vitreous; white, occasionally pale shades of gray, red, green, blue, violet, yellow, brown; embraces variety called Iceland spar; a very common mineral; the essential basis of limestone.

Chalcedony. A crypto-crystalline variety of quartz having a wax-like luster, either transparent or translucent; white, grayish, pale brown to dark brown, black, sometimes delicate blue, occasionally other shades; frequently occurs as the lining or filling of cavities, taking on a botryoidal or mamillary form.

Chlorite. The type of an important group of secondary minerals usually characterized by a green color, softness, and smoothness or unctuousness of feeling; they are usually aluminum-magnesium-iron silicates, with chemically combined water; derived from several other species, as pyroxene, amphibole, biotite, garnet, etc.; embraces a number of species, among which are clinochlore, penninite, prochlorite, and delessite.

Chrysolite. Olivine; essentially a magnesium-iron silicate; orthorhombic; H. 6–7; luster vitreous; green, commonly olive-green, sometimes yellow, brownish, grayish green; highly infusible; a common constituent of certain basic igneous rocks; the name olivine is more commonly used by geologists.

Corundum. Alumina; oxide of aluminum; H. 9; rhombohedral; large crystals usually rough; luster vitreous; color blue, red, yellow, gray, and nearly white; the bluish variety is *sapphire;* the red variety is *ruby*, the yellow, oriental *topaz*, the green, oriental *emerald*, and the purple oriental *amethyst;* dark colors, with iron oxide, *emery.*

Diallage. A variety of pyroxene; H. 4; characterized by thin foliae; usually grayish green to grass-green, or deep green; luster on cleavage surface pearly, sometimes metalloid or brassy; an essential mineral in the gabbros, as sometimes defined.

Enstatite. One of the pyroxenes; essentially a magnesium silicate; orthorhombic; H. 5.5; luster a little pearly on cleavage surface; metalloidal in the bronze variety (bronzite); grayish white, yellowish white; greenish white to olive-green and brown; very infusible; a common mineral in certain basic crystalline rocks.

Epidote. A complex aluminum-calcium-iron silicate of varying composition; monoclinic; H. 6–7; luster vitreous, pearly, or resinous; color usually pistachio-green, or yellowish green to brownish green; common in many crystalline rocks, usually as a secondary product.

Feldspars. Aluminum silicates, with either potassium, sodium, or calcium, or two or more of these; crystal systems, monoclinic and triclinic; possess distinct cleavage in two directions; H. 6–6.5; range in color from white through pale yellow, red, or green; and occasionally dark; triclinic feldspars frequently called *plagioclase;* the group embraces orthoclase, microcline, albite, oligoclase, andesine, labradorite, anorthite, and numerous varieties.

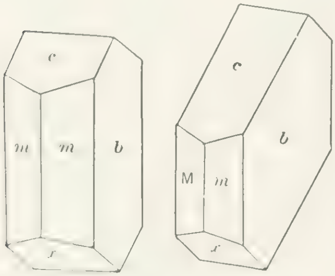

Fig. 329. — Feldspar crystals.

Fluorite. Fluorspar; calcium fluoride; isometric, usually cubic; H. 4; luster vitreous, sometimes splendent; white, yellow, green, rose, crimson-red, violet, sky-blue, and brown; yellow, greenish, and violet most common; occurs usually in veins or cavities in beautiful crystalline form.

Galena. Lead sulphide; isometric, usually cubic; perfect cubic cleavage; luster metallic; lead-gray; a common ore of lead; occurs in veins and layers, also as linings of cavities.

Garnet. A complex silicate of varying composition, embracing aluminum, calcium, magnesium, chromium, iron, and manganese, but usually only two or three of these are present in abundance, and each variety is characterized by its leading constituent; isometric, usually in dodecahedrons or trapezohedrons; H. 6.5–7.5; luster vitreous to resinous; commonly red or brown, sometimes yellow, white to blue, green or black; common in mica schist, gneiss, hornblende schist; also in granite, syenite, and metamorphosed limestone.

Glauconite. Green-sand, a hydrous potassium-iron silicate usually impure, amorphous, or earthy; dull olive-green or blackish, yellowish, or grayish green; opaque, commonly occurs as grains or small aggregations.

Graphite. Plumbago, black lead; a form of carbon, usually impure; rhombohedral, but rarely appearing as a crystal; more often as thin laminae of greasy feel; yields a black adhesive powder; hence its common use for lead pencils; occurs in granite, gneiss, mica schist, crystalline limestone; sometimes results from alteration of coal by heat; occasionally occurs in basaltic rocks and meteorites.

Gypsum. A hydrous calcium sulphate; monoclinic; perfect cleavage into plates; occurs in a variety of forms, including fibrous and granular; H. 1.5–2; luster pearly and shiny; white, sometimes gray, flesh-red, yellowish, and blue; impure varieties dark; crystallized varieties include selenite, satin-spar, alabaster, etc.; easily recognized by its softness and lack of effervescence with acids; occurs in beds; calcined and ground to make plaster of paris.

Hematite. Ferric oxide, Fe_2O_3; rhombohedral, more commonly columnar, granular, botryoidal, or stalactitic; luster metallic, sometimes earthy; iron-black, dark steel-gray, red when earthy; gives red streak or powder; a leading iron ore, 70 percent metallic iron when pure; the chief source of the red color of soils and rocks generally.

Hornblende. The most important amphibole; name sometimes used as a synonym for amphibole.

Hypersthene. One of the pyroxenes; a ferromagnesian silicate; orthorhombic; H. 5–6; luster somewhat pearly on cleavage; surface often iridescent; dark brownish green, grayish, or greenish black and brown; a frequent constituent of crystalline rocks.

Iceland Spar. A form of transparent calcite.

Ilmenite. A titanium iron-oxide; rhombohedral; resembles hematite; luster submetallic; iron-black; powder black or brownish red; occurs frequently in crystalline rocks associated with magnetite.

Iron Pyrites. Pyrite.

Kaolin. Kaolinite; essentially a hydrous aluminum silicate; usually in clay-like or earthy form; white or grayish white; often tinged with impurities; commonly arises from decomposition of aluminous silicates, especially the feldspars; basis of pottery and china.

Labradorite. A plagioclase feldspar; essentially an aluminum-calcium-sodium silicate; composition intermediate between that of albite and anorthite; triclinic; H. 6; luster pearly or vitreous, gray, brown, or greenish; sometimes colorless or white; frequently shows play of colors; important constituent of various crystalline rocks, especially of the basic class; usually associated with a pyroxene or amphibole.

Leucite. Essentially an aluminum-potassium silicate, allied to the feldspars; H. 5–6; luster vitreous, white, ash-gray, or smoke-gray occurs in certain volcanic rocks, particularly lavas of Vesuvius.

Limonite. Ocher; a hydrous iron oxide; commonly earthy; also concretionary, stalactitic, botryoidal, and mamillary, with fibrous structure; H. 5–5.5; luster silky, sometimes submetallic, but commonly dull and earthy; brown, ocherous yellow; streak and powder yellowish brown; constitutes ocher, bog-ore, ironstone, etc.; is the chief source of the yellow color of soils and rocks; arises from the alteration of other iron ores.

Magnesite. Magnesium carbonate; rhombohedral; white, yellowish, grayish white to brown; fibrous, earthy, or massive; found in altered magnesium rocks.

Magnetite. Magnetic iron ore; iron oxide, Fe_3O_4; octahedral or dodecahedral; strongly magnetic; H. 5.5–6.5; abounds in igneous and metamorphic rocks.

Marcasite. White iron pyrites; iron sulphide; same composition as pyrite, which it closely resembles; H. 6–6.5; luster metallic, pale gray, bronze, or yellow; prone to decomposition; disseminated through various rocks, particularly plastic clays containing organic matter.

Mica. The type of an important group of rock-forming minerals well known for their perfect cleavage into thin elastic laminae; among the leading varieties are the common potassium mica (muscovite), the sodium mica (paragonite), the lithium mica (lepidolite), the magnesium-iron mica (biotite), the magnesium mica (phlogopite), and the iron-potash mica (lepidomelane).

Microcline. A triclinic feldspar, closely resembling orthoclase in appearance and having the same composition.

Muscovite. Common or potash mica; essentially an aluminum-potassium silicate; H. 2–2.5; monoclinic; remarkable for its basal cleavage; splits easily into exceedingly thin, flexible, elastic laminae; luster vitreous, more or less pearly or silky; colorless or variously tinged brown, green, or violet; a common mineral in crystalline rocks, particularly in the granites, gneisses or schists.

Oligoclase. A plagioclase feldspar; essentially an aluminum-calcium-sodium silicate which may be regarded as a mixture of albite and a small amount of anorthite; triclinic; luster vitreous, pearly, or waxy; whitish grading into greenish and reddish; H. 6–7; common in crystalline rocks.

Orthoclase. A potash feldspar; essentially a potassium-aluminum silicate; varying by the replacement of the potassium by sodium and less frequently by other substitutions; monoclinic; occurring in distinct crystals and also in crypto-crystalline forms; cleavage planes perfect with pearly luster on cleavage surface; white, gray, and flesh-red, occasionally varying to greenish white and bright green; H. 6–6.5; difficultly fusible; sanidine a glassy variety; a very common mineral, especially in the granites, syenites, and gneisses.

Olivine. Chrysolite.

Petroleum. A native mineral oil; a mutual solution of hydrocarbons, commonly believed to arise from organic matter, both animal and vegetable.

Plagioclase. A general term embracing the triclinic feldspars whose two cleavages are oblique to each other; embracing albite, oligoclase, andesine, labradorite, and anorthite.

Plumbago. Graphite.

Pyrite. Iron pyrites, fool's gold, iron disulphide; isometric; commonly in cubes; H. 6.5; luster metallic, splendent, or glistening; pale brass-yellow; occurs widely disseminated throughout a large class of rocks; usually harder and lighter in color than copper pyrites, and deeper in color than marcasite which has the same composition.

Pyroxene. The type of a large and important group of rock-forming ferromagnesian minerals; varies in composition and embraces a large number of varieties; usually a magnesium-iron-calcium silicate; crystals usually thick and stout, but varying greatly; sometimes lamellar and fibrous; H. 5–6; luster vitreous inclining to resinous; green of various shades verging occasionally towards light colors, more often to browns and blacks; among the minerals belonging to the pyroxene group are augite, bronzite, diallage, diopside, enstatite, hypersthene, and others.

Quartz. Crystallized silica; hexagonal; crystals commonly six-sided prisms capped by six-sided pyramids; without cleavage; H. 7; hard enough to scratch glass; usually transparent, glassy, colorless when pure; shaded by impurities to yellow, red, brown, green, blue, and black; varieties, *amethyst*, purple or violet; *false topaz*, yellow; *rose-quartz*, *smoky quartz*, *milky quartz*, etc.; *chalcedony* is a crypto-crystalline (crystals not visible) variety; *carnelian*, a red chalcedony; *agate*, a variegated or banded chalcedony; *moss-agate*, a chalcedony containing moss-like or dendritic crystallizations of iron or manganese oxide; *onyx*, chalcedony in layers; *jasper*, an opaque, colored quartz, usually red or brown; *flint*, an opaque impure chalcedony; *chert*, an ill-defined term applied to a whitish flinty rock; *hornstone*, a translucent, brittle, flinty rock.

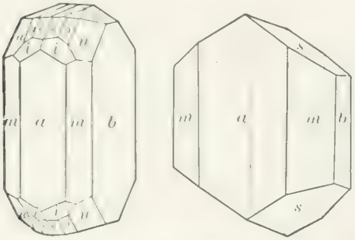

Fig. 330. — Pyroxene crystals.

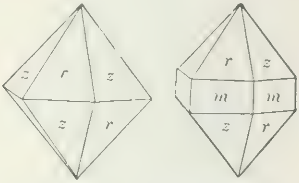

Fig. 331. — Quartz crystals.

Serpentine. A hydrous magnesium silicate; usually in pseudomorphic forms; also fibrous, granular, cryptocrystalline, and amorphous; H. 2.5–4; luster subresinous to greasy, pearly or earthy, resinous or waxlike; feel, smooth and somewhat greasy; leek-green to blackish green and siskin green verging into brownish and other colors; apparently derived most commonly from chrysolite or olivine, but also from other magnesian minerals; sometimes constitutes the bulk of rock masses.

Siderite. Iron carbonate; rhombohedral; H. 3.5–4.5; luster vitreous, more or less pearly, ash-gray, yellowish or greenish, also brownish; occurs as extensive iron deposits and in crystalline rocks.

Steatite. Soapstone, a variety of talc; hydrous magnesium silicate.

Sulphur. A well-known element occurring native in volcanic regions; also formed by the decomposition of sulphides, particularly pyrites.

Talc. A hydrous magnesium silicate; usually in foliae; granular or fibrous forms; also compact; easy cleavage into thin flexible laminae, but not elastic; luster pearly on cleavage surface; apple-green to silvery white; H. 1–2; a secondary product from the alteration of magnesium minerals; distinguished by its soft, soapy feel, soapstone being one variety; whitish form is known as French chalk.

Topaz. An aluminum silicate, with part of the oxygen replaced by fluorine; orthorhombic; H. 8; uster vitreous; colorless, straw-yellow verging to various pale shades, grayish, greenish, bluish, and reddish;

distinguished by its hardness and infusibility; occurs in crystalline rock.

Tremolite. A calcium-magnesium amphibole; a common constituent of certain crystalline rocks.

Zeolites. A group of minerals derived from the alteration of various aluminous silicates.

Classification of Igneous Rocks

Igneous rocks are best classified upon the basis of (1) their *texture*, which expresses in general the conditions of solidification, and (2) their *mineral composition*, which expresses in general the chemical composition. The following table gives a practical classification which can be used without special laboratory methods.

Texture	Color.			
	Light Colored.		Dark Colored.	
Pyroclastic.	Tuff, agglomerate.			
Non-crystalline glassy.	Obsidian, pitchstone, pumice.			
Fine-grained or dense.	Felsite Group.		Basalt Group.	
Coarse crystals in fine-grained groundmass.	Felsite Porphyry Group.		Basalt Porphyry Group.	
Coarse-grained.	With Quartz.	Without Quartz.	With Feldspar.	Without Feldspar.
	Granite Group.	Syenite Group.	Diorite, Gabbro Group.	Perknite Group.

Fig. 332. — Field classification of igneous rocks.

The ultimate classification of igneous rocks is based upon more elaborate determinations of the constituent minerals with the aid of the polarizing microscope, and is only for those having special training in petrography.

The outstanding features involved in the current classification of igneous rocks have been noted in other connections, but they may be recapitulated here. All fragmental igneous rocks are *pyroclastic*, and pyroclastic rocks may be *tuffs, agglomerates*, etc. (p. 350). Rock formed by solidification of lava without the development of crystals, is *obsidian*, if not porous. If porous (hardened rock froth), it is *pumice, scoriaceous glass*, etc. If the rock is composed largely of glass but partly of small crystals, it is sometimes called *pitchstone*, because its freshly fractured surface looks like pitch or resin. When scoriaceous rock has its cavities filled by mineral-matter deposited from solution, it becomes an *amygdaloid*. *Porphyry* has already been defined (p. 357). All these names are based on texture, rather than on mineralogical or chemical composition.

While most igneous rocks are wholly crystalline, and are classified on the basis of their mineralogical composition, their mineral content in turn is determined by their chemical composition. It will be seen, therefore, that the chemical composition is the more fundamental. The number of varieties of igneous rocks is very large, but only a few of the more important ones need be mentioned here.

The Granites. This name was originally used to designate a granular, megascopically crystalline rock and it is still popularly and properly so used. In scientific treatises it has usually been confined to rock composed chiefly of crystals of quartz, feldspar (predominantly orthoclase, p. 367), and mica. To satisfy the need for a useful general field term the word "granite" has been used to include again all coarse-grained igneous rocks composed chiefly of quartz, feldspar of any kind, with mica, hornblende, or other minerals in subordinate amounts. In normal granite, the crystals are distinct and in some cases large (Fig. 325). They are intimately mingled with one another, and in growing, have interfered more or less with one another and so became interlocked. The granites are among the most common and easily recognized of the coarse-grained igneous rocks. Their color is determined largely by the feldspar, the red and pink varieties of the mineral giving rise to red granite, and the whitish varieties to gray granite.

Granites vary widely from their type form by the addition and substitution of other minerals, and these sometimes become as prominent as the type minerals. Thus other feldspars sometimes take the place of the orthoclase, or accompany it; hornblende and other minerals take the place of the biotite, or occur with it, and so on. Whenever one of these replacing or accessory minerals is abundant, its name is often prefixed, as hornblende-granite. Granite grades almost insensibly into other types of igneous rock, as syenite, diorite, etc.

The granites were formed from a lava rich in silica (68–70 percent), alumina, potash and soda, but generally poor in lime, iron, and magnesia. Incidentally other substances are present. Granite is generally an intrusive massive rock which solidified slowly under conditions which permitted complete crystallization, with the development of large crystals.

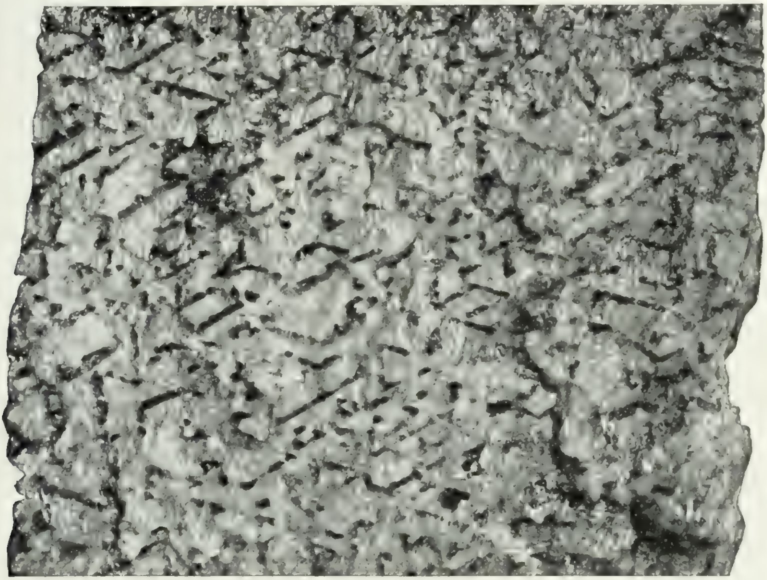

Fig. 333. — Graphic granitic (or pegmatitic) texture. Nearly natural size.; (Church.)

When rock of the composition of granite is finely banded, it is *gneiss*. The texture of *graphic granite* (see pegmatite, p. 23) is notably peculiar, due to the simultaneous crystallization of the quartz and feldspar (Fig. 333).

The Syenites. The term syenite (from Syene on the Nile) is now applied to a rock consisting essentially of feldspar, with mica or hornblende, or both, which contains little or no quartz; but there is a complete gradation from the granites to the syenites. The syenites are richer in iron and magnesium than the granites, and poorer in silica (about 58–60%). The syenites also grade into other classes of rock as do the granites, and are named by similar prefixes, as augite-syenite, etc. The syenites are red or gray according to the color of the feldspar. The texture of syenite is like that of granite.

The Diorites. The diorites are rocks which crystallized from lavas having about the same amount of silica as the lavas of the syenites, but poorer in soda and potash and richer in lime and magnesia. In current usage, diorite is defined as a rock composed of an intimate mixture of crystals of hornblende and a plagioclase feldspar (p. 367). It differs from syenite in having plagioclase feldspar instead of orthoclase. By substitutions and additions of accessory minerals, the diorites grade toward the granites and syenites on the one hand, and toward the gabbros on the other.

The Gabbros. The name gabbro was formerly applied to a coarse-grained basic rock consisting of labradorite (p. 366) and diallage (p. 364) but the name has been extended gradually until it embraces a large group of rocks whose principal minerals are plagioclase and pyroxene. The gabbros are mostly dark colored and rather heavy.

The Perknites (peridotites and pyroxenites). These stand at the basic end of the series, having been formed from a magma in which the silica content was low (39–45 percent), as were also alumina, lime, soda, and potash, but in which the magnesia content was high (35–48 percent). The peridotites consist very largely of the mineral *olivine* (a magnesium iron silicate) associated with pyroxene, magnetite, and other very basic minerals. Little or no feldspar is present. They are much less abundant than the preceding classes and usually occur as dikes.

Closely allied to the peridotites are rocks which are made up largely of a single basic mineral, as *pyroxenite*, *hornblendite*, rocks essentially formed of the minerals pyroxene and hornblende respectively.

The Felsites. The term felsite is used in a comprehensive way for light-colored, compact igneous rocks. They are the fine-grained equivalents of the coarse granites and syenites. Quartz and the feldspathic minerals predominate over ferromagnesian minerals. Rhyolite, dacite, trachyte and light-colored andesites are lavas belonging to the felsite group. They are distinguished by the kinds of feldspar composing them. They are classed as acidic lavas, being richer in silica than the basalts.

The Basalts. The term basalt is used in a somewhat comprehensive way for dark, compact igneous rocks that appear to be nearly homogeneous, owing to the smallness of the crystals, which are usually so minute as to be identifiable only under a microscope. The basalts are the fine-grained equivalents of the coarser gabbros and perknites. The leading minerals are plagioclase (usually labradorite) and pyroxene (usually augite), with or without olivine, and magnetite or ilmenite present. There is a considerable range in chemical composition, but the basalts are relatively poor in silica and usually in potash and soda, but rich in lime, magnesia and iron. The ferromagnesian minerals predominate over the feldspathic minerals. The basalts are classed as basic, and some are highly so. The lavas of basaltic flows were especially fluid, and when poured out upon the surface, spread out in thin sheets.

General Names. The difficulty of distinguishing many of the foregoing rocks from one another by any means available in the field, owing to the minuteness of the crystals, and to the gradation of one type of rock into another, makes it desirable to employ certain general names which will correctly express the leading character of the rock without implying a knowledge of the precise mineral composition. A convenient term for any dark rock is *greenstone*, which merely indicates that the ferromagnesian minerals are predominant and usually give a greenish or greenish black cast to the rock. The greenstones embrace some of the diorites, gabbros, and the basalts, and may even extend to the peridotites and perhaps to others. Another convenient name is *trap*, which may be used for any dark, fine-grained igneous rock, such as basalt. The term basalt itself is used in much the same way.

Varieties of Rock Dependent upon Conditions

From what has preceded, it is clear that the chemical nature of the liquid magma determines the mineralogical composition of the rock, if it is crystalline. But it may now be pointed out that the same magma which made a plutonic granite, might have made a porphyry, a felsite, an obsidian, a pumice, or a tuff, under other conditions of cooling and hardening. The same is true of other varieties of coarse-grained rocks.

CAUSES OF VULCANISM

The explanation of the extraordinary facts involved in volcanic phenomena is wrapped up in that of the origin of the earth, for the agencies which the earth inherited from its birth are beyond doubt factors in vulcanism. This phase of the subject will be treated here only very briefly. Let us consider first some of the factors which are involved.

Fluxing. (1) The great body of the earth is thought of as made of heterogeneous material, mixed as it accumulated on the surface of the growing earth and buried successively by further accretion. Gases and substances which produce gases when heated, were incorporated in the body of the solid earth-material as it accumulated.

(2) As pointed out in the discussion of diastrophism, pressure acting upon the heterogeneous material of the earth produces compression and generates heat. The temperature of the body of the earth therefore rises. This rising temperature results in —

(3) The solution at certain spots of the most soluble materials in each other, i.e., *selective fluxing* and the formation of lava. The presence of radio-active substances may help to locate the fluxing by raising the temperature in their immediate vicinity above that normal to the zone in which they lie. Thus, while the great mass of the earth remains solid, small pockets of lava are here and there generated.

(4) The liquid portions, which are lighter than the solid rock, are caused to rise by gravitative pressure assisted by the kneading effects of differential stresses, such as diastrophic, tidal, nutational, and rotational forces. As the small masses of lava rise and enter zones of less pressure, they dissolve additional matter. This is because they have, on account of the diminished pressure, more heat than is necessary to keep them in a liquid condition. This overhead fluxing is doubtless assisted by the presence of more abundant radio-active material and by the concentration of gases at the top of the rising column. It is also part of this concept that, as the lava threads rise into higher and cooler levels of less pressure by absorbing overhead material, they precipitate out and leave behind material insoluble under the changed conditions. Thus it is conceived that these fluxing selections thread their way outward through the solid earth, till some of them reach the zone of fracture, and thence, taking advantage of fissures and fractures, reach the surface (Fig. 334).

The ascending column may solidify before going far; it may solidify near the surface, to form the various sorts of intrusive structures; or it may find equilibrium only by coming to the surface and pouring out a portion of its substance and discharging its gases. The majority of lava threads solidify before reaching the surface, but in their ascent they have been a means of bringing heat outward from the deep interior, raising the temperature of the outer part, and thus making the way easier for other lavas to follow.

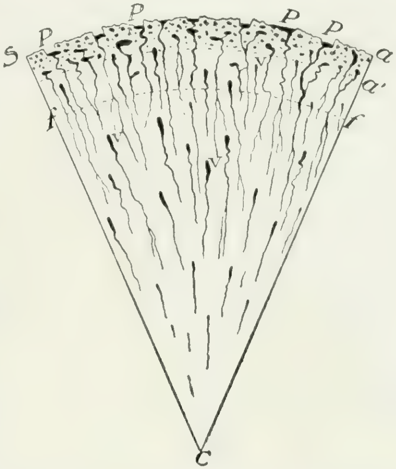

Fig. 334. — Ideal section of a portion of the early earth, illustrating its assigned modes of vulcanism. *C*, center; *S*, surface; *a–a*, fragmental zone; *a–f*, zone of continuous rock below surface melting temperature; *ff–c*, interior portion whose temperatures rise from the surface melting-point at *f–f* to a maximum at *C;* *V*, *V*, threads or tongues of molten rocks rising from the interior to various levels, many of these lodging within the fragmental zone as tongues, batholiths, etc.; *PPP*, explosion pits formed by volcanic gases derived from tongues of lava below.

Relief of Pressure. The interior of the earth is at a high temperature, though various lines of evidence show that it is solid except for very limited, temporarily liquid, threads of lava. Measurements of temperature in deep mines and wells show an average increase of 1° C. for every 100 to 125 feet of descent. If this rate of increase continues unchanged with increasing depth, the temperature at a depth of 60 miles would be over 3000° C. There are, however, reasons for believing that 1500° C. is nearer the truth. In any case, the temperature is sufficiently high to liquefy rocks at the surface. But at such depths, the pressure is so tremendous that the rocks as a whole are prevented from liquefying, and are held solid in spite of the high temperature. Any reduction of pressure, therefore, facilitates liquefaction, and this must be an important factor in vulcanism. Partial relief of pressure occurs locally in certain types of mountain building (beneath strata which are arched up by lateral pressure) and as an accompaniment of other phases of diastrophism. Such partial relief of pressure may be sufficient to cause small selected portions of the solid rock to become liquid lava, and also to aid in the movement of any lava however it may have been generated. Thus intrusions of granite and related rocks have been particularly abundant beneath the most strongly up-arched portions of folded mountain ranges where, after erosion, the granite, etc., becomes exposed as the "core" of the range.

The general solidity of the earth is maintained under the conditions of high pressure by the liquefaction and removal of the most soluble parts, while at the same time vulcanism acts as a safety valve to let out much of the accumulated heat. The independence of volcanoes is assigned to the independence of the liquid threads that work their way to the surface. Nothing like a great reservoir or molten lake enters into the conception. Sympathetic action between volcanoes is attributed to a common expelling force acting on the liquid threads. The distribution of volcanoes, past and present, in essentially all latitudes and longitudes, is assigned to the general all-pervading nature of the cause. The special surface distribution in belts, zones, and lines is, in part, attributed to, though not altogether controlled by, relief of pressure resulting from diastrophic movements and in part to favorable conditions of escape in these places. Thus volcanoes are to be seen in the vicinity of the zones of weakness developed between the great segments of the earth where, in general, more than an average amount of lava is generated, and where escape is easiest.

CONCLUSION

Because of the very nature of vulcanism, many of its phases are wrapped in mystery, and their explanations are necessarily hypothetical. This is also true, to a considerable extent, of diastrophism, the causes of glaciation, the descent of ground-water and manifestly of atmospheric circulation. The explanatory hypotheses are based upon a large group of very definite and significant facts. These hypotheses, while representing the best opinion at the present time, are open to modification whenever new facts bring additional light to bear. In the meantime they are useful, not only in giving concrete conceptions of geologic processes, but what is more important, they direct the search for those particular facts which are most needed for further advancement of the subject. Hypotheses are the greatest stimulants to thought and research.

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